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THESIS

COMPARISON OF MEASURED AND TRANSFORMED DIRECTIONAL WAVE SPECTRA USING LINEAR REFRACTION MODEL

by

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Comparison of Measured and Transformed Directional Wave Spectra Using Linear Refraction Model

by

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ABSTRACT

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TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

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I dedicate this thesis to my wife, Farida, and children Salman, Talia and Usman.

I. INTRODUCTION

Wave climatology plays an important role in coastal dynamics as most energy is put into coastal region by wind-generated waves. Initially, ocean wave analysis was limited to a single wave train defined by a wave height and a period in a one-dimensional spectrum, but in the last couple of decades the emphasis has been placed on two-dimensional spectra.

The shoaling waves undergo refraction, reflection and diffraction mostly influenced by the wave approach angle and the bathymetry of the area. A spectral approach requires the superposition of a large number of progressive sinusoidal component waves. Therefore, waves of small amplitude (linear wave) are assumed. This assumption neglects reflection, diffraction and postulates long-crested waves.

The area of Monterey Bay has been selected for this comparative study of ocean waves. This area presents an interesting problem for predicting directional wave spectra because of its variable shoreline orientation relative to prevalent eastward waves and the presence of the Monterey submarine canyon bisecting the bay. The Monterey submarine canyon is the largest in the western hemisphere and causes substantial perturbations on the nearshore wave field.

Historically, Monterey Bay was one of the first, and presumably the most intensively studied, locations for wave refraction. Wave refraction diagrams of Monterey Bay were first constructed in 1948 by Johnson, O'Brien and Isaac (Wiegel, 1964). The first intensive refraction studies were performed for amphibious landing exercises at Fort Ord (1954) followed by studies by the Corps of Engineers (1958) for small craft harbor at Santa Cruz. Wave refraction and littoral drift calculations have been performed by Dorman (1969) and Arnal et al. (1973). All these studies were performed using hand-drawn refraction diagrams. More recently, numerical refraction calculations have been made for a number of specific marine construction sites (e.g., Thornton (1978, 1980)) and erosion studies (Oradiwe, 1986; McGee, 1986).

The objective of this study is to compare ocean wave measurements with the linear refraction model. The deep water two-dimensional spectrum is transformed to shallow water spectrum using transfer function computed from a linear wave refraction model by Dobson (1967). The deep water spectrum, $S_0(f_0, \theta_0)$ is determined from data acquired from the National Data Buoy Center (NDBC) using the Fourier analysis technique as proposed by Longuet-Higgins (1963) and Cartwright (1963). The transformed spectrum $S(f, \theta)$, is compared with measured shallow wave spectra for which the

data had been acquired from pressure gauge arrays located at Marina and Santa Cruz.

The background theory is described in Chapter II. The data acquisition and equipment are mentioned in Chapter III. Chapters IV and V contain discussions and the conclusions.

II. THEORETICAL BACKGROUND

A. WAVE REFRACTION

1. General

The refraction of ocean surface waves on the continental shelf is necessary for forecasting ocean waves in shallow water. Practical techniques for calculating refraction diagrams were developed by Johnson (1948) and Munk and Arthur (1952). The graphical technique developed by these investigators is called the Orthogonal Technique. With the advance of digital computers, the numerical methods were introduced by Griswold (1963), Wilson (1966) and Dobson (1967). The numerical calculations of wave orthogonals (or rays) can be quickly and accurately plotted for engineering purposes. The study of wave transformation that includes refraction over complex underwater topography and in large areas is more commonly accomplished through numerical models. (Hayes, 1980)

2. Refraction Theory

In this study, linear wave theory is applied to shoaling and refracting waves, which neglects reflection and diffraction assuming energy propagates only in the direction in which the waves travel. It also neglects amplitude dispersion. The wave refraction is based on generalization of Snell's law for phase speed variation and conservation of

energy. The phase speed of the wave depends upon wave period (or frequency) and the depth of water at particular locations. Wave energy is conserved between wave orthogonals, where orthogonal is perpendicular to the wave crest. The orthogonal indicates the direction in which shoaling wave progresses. Based on linear wave theory, it is assumed that the wave period will remain constant but the wavelength, the velocity and the height, will vary as the wave shoals toward the beach.

In general, the phase speed of a small-amplitude linear water wave is expressed as

$$C^2 = \frac{h}{k} \tanh(kd) \tag{2.1}$$

where g is acceleration of gravity, k the wave number and d is the depth of water at a particular location. Consequently, in shallow water the phase speed of the wave varies with depth only and Snell's law becomes applicable in the case of straight and parallel contours as

$$\frac{\sin}{\sin_{\quad 0}} = \frac{C}{C_0} \tag{2.2}$$

where is an angle between the wave crest and local depth contours and C is phase speed. The subscript zero refers to deep water conditions.

The energy of a wave when it reaches the beach is a function of the wave height, and the wave height at the shore is related to the wave height in deep water. The variation of wave height within two wave rays is described (Mei, 1980) as

$$H = H_0 [\frac{Cg_0}{Cg}]^{1/2} [\frac{b_0}{b}]^{1/2} \tag{2.3}$$

or

$$H = H_0 \, K_S \, K_R \tag{2.4}$$

where

$$K_S = [\frac{Cg_0}{Cg}]^{1/2} = \text{Shoaling Coefficient} \tag{2.5}$$

$$K_R = [\frac{b_0}{b}]^{1/2} = \text{Refraction Coefficient} \tag{2.6}$$

where b is the distance between two rays and Cg is group velocity and may also be represented in the form of

$$C_g = \frac{C}{2} [1 + \frac{2\,kd}{\sinh\, 2kd}]$$

Using the relationship $C = C_0 \tanh(kd)$ (Kinsman, 1965), Equation (2.5) may be written as

$$K_S = [\frac{k}{k_0}]^{1/2} [1 + \frac{2\ kd}{\sinh\ 2kd}]^{-1/2} \tag{2.5a}$$

From Equation (2.2), the change in angle determines the change in wave crest or b value. Thus the value of K_R may be computed from

$$\frac{b_0}{\cos \alpha_0} = \frac{b}{\cos \alpha}$$

or

$$K_R = [\frac{b_0}{b}]^{1/2} = [\frac{\cos \alpha_0}{\cos \alpha}]^{1/2} \tag{2.6a}$$

where

$$= \sin^{-1}[C \frac{\sin \alpha_0}{C_0}]$$

In any linear wave refraction model, shoaling and refraction coefficients are first computed and then the wave height is determined.

3. Linear Wave Refraction Model

Munk and Arthur (1952) described the theory of wave intensity along a refracted wave. Dobson (1967) followed the theory and developed a linear refraction model. The Dobson model is briefly described here.

The definition sketch for the wave ray equation is shown in Figure 2.1 in that $\overline{A_1B_1}$ is tangential to circular

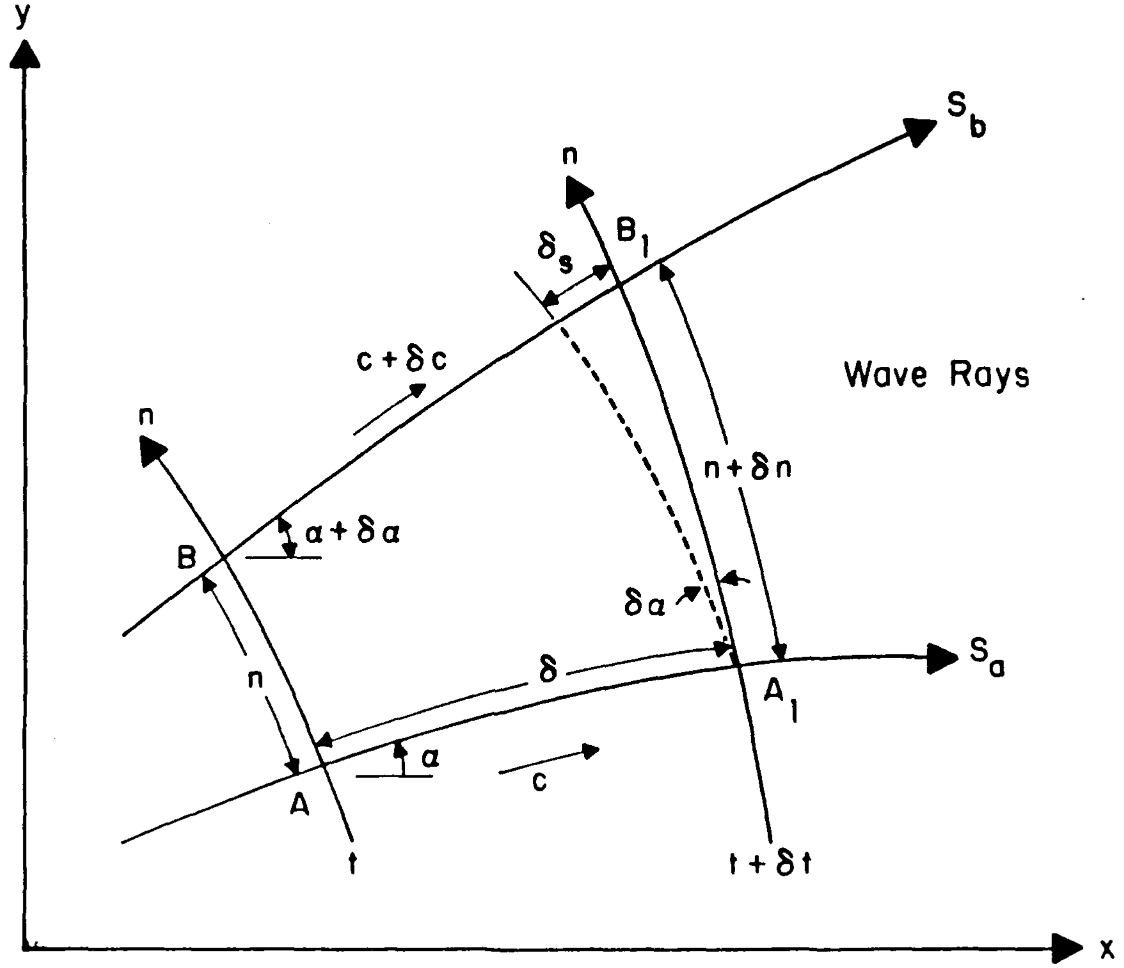


Figure 2.1 Definition Sketch for Wave Ray Equations (Dobson, 1967)

arcs of radius $C\delta t$ and $(C+\delta C)\delta t$ and S and n are very small. Two basic characteristic equations for the propagation of a wave ray are the ray curvature equation

$$R = \frac{D\alpha}{DS} = -\frac{1}{C}\frac{DC}{Dn} \tag{2.8}$$

and an equation for the ray separation factor β (Dobson, 1967)

$$\frac{d\beta^2}{dt^2} + \frac{d\beta}{dt}\frac{DC}{DS} + \beta C\frac{D^2C}{Dn^2} = 0 \tag{2.9}$$

Dobson manipulated these equations using Gauss integral theory into a convenient form where Eqs. (2.8) and (2.9) can be easily applied in terms of cartesian coordinates and time variable as

$$R = \frac{1}{C}[\sin\alpha\frac{\partial C}{\partial x} - \cos\alpha\frac{\partial C}{\partial y}] \tag{2.10}$$

$$\frac{d^2\beta}{dt^2} + p\frac{d\beta}{dt} + q\beta = 0 \tag{2.11}$$

where

$$P(t) = 2[\cos\alpha\frac{\partial C}{\partial x} + \sin\alpha\frac{\partial C}{\partial y}] \tag{2.12}$$

$$q(t) = C[\sin^2\alpha\frac{\partial^2 C}{\partial x^2} - 2\sin\alpha\cos\alpha\frac{\partial^2 C}{\partial x\partial y} + \cos^2\alpha\frac{\partial^2 C}{\partial y^2}] \tag{2.13}$$

The phase speed is a function of the coordinates so that $C = C(x,y)$. The wave ray can be constructed with Equation (2.10) within a coordinate system. The ray curvature R is computed using finite difference methods between two consecutive points along a wave ray by iterating

$$\Delta\alpha = \bar{R}\,\Delta S \tag{2.14}$$

at each step. So the new coordinates can be computed as

$$x_{S+\Delta S} = x_S + \Delta S \cos \bar{\alpha} \tag{2.15}$$

$$y_{S+\Delta S} = y_S + \Delta S \cos \bar{\alpha} \tag{2.16}$$

where:

$$\bar{R} = \frac{R_S + R_{S+\Delta S}}{2}$$

$$\bar{\alpha} = \frac{\alpha_S + \alpha_{S+\Delta S}}{2}$$

The ray path is constructed grid to grid over the sea bottom until it reaches the required location.

The ray separation factor β in Equation (2.11) is computed applying Fox's formula for a second order linear equation. This is a finite difference method and requires that the separation factor be known at the starting and

preceding points. This condition is conveniently satisfied by starting the ray in the deep water where is equal to 1.

The characteristics of the wave at any point in the coordinate system may be determined from the solution of Equations (2.10) and (2.11) provided the speed, $C(x,y)$ is known. However it is common to consider waves of different deep water characteristics travelling towards the coastline over a known bathymetry, so that it is more convenient to consider the problem in terms of water depth, $d(x,y)$ rather than wave speed. Thus these equations are expressed as functions of the mean water depth.

Dobson used a quadratic least square surface to fit depth locally at intermediate points using

$$d = {}_1 + {}_2x + {}_3y + {}_4x^2 + {}_5xy + {}_6y^2 \tag{2.17}$$

where the partial derivative of depth, d, with respect to x and y is expressed as functions of . Therefore the local water depth can be obtained by interpolation method.

The bathymetry used in this refraction program was obtained from the original NOAA data and projected onto a six second (latitude and longitude grid) modified Universal Transverse Mercator (UTM) grid after screening the data for errors and bad points. Insufficient data were available for off-shore bathymetry. Thus it required interpolation to fill the empty grid cells. Piece-wise linear triangulation

was used to interpolate to a 200 meter rectangular grid. A nine-point weighted linear average was applied to smooth bathymetry (Oradiwe, 1986; McGee, 1986).

B. DIRECTIONAL WAVE SPECTRUM

1. General

Directional characteristics of ocean waves are important, not only for understanding wave generation mechanisms but also for practical problems such as wave forecasts in the design of coast and harbor protection works. Directionality is equally important in the refraction of ocean waves. Various techniques incorporating floating buoys, pressure gauge arrays, electromagnetic current meters and SAR have been utilized to determine directional properties of the ocean waves. In this study data have been acquired from the NDBC heave/pitch/roll buoy for deep water spectra and pressure gauge arrays for shallow water. The theoretical background of these measurements is presented in the following sections.

2. Deep Water Wave Spectrum

Longuet-Higgins (1963) suggested a method to compute directional wave spectrum from water surface elevation and two orthogonal components of wave surface slope. The surface elevation is supposed to be the result of the superposition of a large number of long-crested progressive sinusoidal component waves in random phase (Battjes, 1972). Several other investigators (e.g., Cartwright, 1963; Long,

1980; Mitsuyasu, 1975; and Long and Hasselmann, 1979) have suggested and refined the technique.

The directional spectrum $S(f,\theta)$ as given by Longuet-Higgins (1963) is expressed in the form

$$S(f,\theta) = S(f) \cdot D(f,\theta) \qquad (2.18)$$

where $S(f)$ is the one-dimensional frequency spectrum and $D(f,\theta)$ is angular distribution function with the condition

$$\int_0^{2\pi} D(f,\theta)d\theta = 1 \qquad (2.19)$$

Equation (2.19) can be expanded by a Fourier series

$$D(f,\theta) = \frac{1}{\pi}\left[\frac{1}{2} + \sum_{n=1}^{\infty}\left(a_n \cos(n\theta) + b_n \sin(n\theta)\right)\right] \qquad (2.20)$$

The coefficients of the first two harmonics can be estimated from the co- and quad-spectra from heave (index 1), and slope signals (indices (1,2), correspond to x,y slopes) (Longuet-Higgins, 1963; Long, 1980) as

$$a_1 = \int_0^{2\pi} D(f,\theta)\cos\theta\, d\theta = \frac{Q_{12}}{K} \qquad (2.21)$$

$$b_1 = \int_0^{2\pi} D(f,\theta)\sin\theta\, d\theta = \frac{Q_{13}}{K} \qquad (2.22)$$

$$a_2 = \int_0^{2\pi} D(f,\alpha) \cos 2\alpha \, d\alpha = \frac{C_{22} - C_{33}}{K^2 C_{11}} \tag{2.23}$$

$$b_2 = \int_0^{2\pi} D(f,\alpha) \sin 2\alpha \, d\alpha = \frac{2C_{33}}{K^2 C_{11}} \tag{2.24}$$

where the constant K is defined as

$$K = \left[\frac{C_{22} + C_{33}}{C_{11}}\right]^{1/2}$$

A heave/pitch/roll buoy and pressure sensor array provide a measure of η, η_x, and η_y at fixed points. The co- and quad-spectra of pairs between these three quantities are computed using the Fast Fourier Transform.

The wave directionalities are determined by the directional distribution function. Equation (2.20) may be rewritten as

$$D(f,\alpha) = \frac{1}{2\pi} + \frac{1}{\pi}[r_1 \cos(\alpha - \alpha_1) + r_2 \cos 2(\alpha - \alpha_2)] \tag{2.25}$$

where

$$r_1 = [a_1^2 + b_1^2]^{1/2} \quad \text{and} \quad r_2 = [a_2^2 + b_2^2]^{1/2}$$

In the case of unimodal spectrum, Longuet-Higgins (1963) proposed a cosine powered distribution function

$$D(f,\alpha) = G(s) \left|\cos \frac{1}{2}(\alpha-\bar{\alpha})\right|^{2S} \quad (2.26)$$

where

$$G(S) = 2^{2S-1} \frac{\Gamma^2(S+1)}{\Gamma(2S+1)} = 2^{2S-1} \frac{(S!)^2}{(2S)!}$$

is a normalizing function to satisfy Equation (2.19). The parameter S is a function of wave frequency and $\bar{\alpha}$ is the mean direction computed from

$$\bar{\alpha} = \tan^{-1}(\frac{b_1}{a_1}) = \tan^{-1}(\frac{Q_{13}}{Q_{12}})$$

Cartwright (1963) showed that the value of S is determined by

$$S_1 = \frac{r_1}{(1 - r_1)}$$

$$S_2 = \frac{1 + 3r_2 + (r_2^2 + 14r_2 + 1)^{1/2}}{2(1 - r_2)}$$

The cosine distribution model is used most in unimodal wave analysis. The same expression is used in this study.

3. Predicted Shallow Water Wave Spectrum

The shallow water wave spectrum can be predicted from the deep water spectrum by multiplying the squares of

both refraction (K_R) and shoaling (K_S) coefficients and the Jacobian of the direction function by each frequency component. Hence the deep water spectrum is transformed into shallow water spectrum by

$$S(f,\theta) = S_0(f_0,\theta_0) K_R^2 \, K_S^2 \, J \tag{2.27}$$

where the Jacobian

$$J = \frac{\partial(f_0,\theta_0)}{\partial(f,\theta)}$$

Since the wave period remains invariant, J only depends upon the changes of local wave direction.

4. Measured Shallow Water Wave Spectrum

The two-dimensional shallow water wave spectrum is determined from the data acquired from shallow water sensors. The same technique as described previously for deep water spectrum is used. In this study, the pressure gauge arrays have been utilized in which subsurface pressure fluctuations are measured to infer the surface elevation (Kim, 1974; Pawka et al., 1976).

From linear wave theory, the relation between wave elevation (η) and subsurface pressure (P) is described by

$$\eta(t) = \left[\frac{\cosh kd}{\cosh k(d+z)}\right] P(t) \tag{2.28}$$

where

d = water depth in meters,

z = vertical distance from mean sea level to pressure gauge (z is positive upward from sea level),

P = pressure measured in meters of water head about a zero-mean pressure.

The term in the brackets in Equation (2.28) is the spectral transfer function relating pressure to surface elevation as a function of frequency and depth. Equation (2.28) can be modified to include differential distance and elevation between two sensors of an array to compute co- and quad-spectra of surface slope as

$$S_{\frac{d\eta}{dx}}(f) = \frac{1}{\Delta x^2}\left[\frac{\cosh kd}{\cosh k(d+z)}\right]^2 S_{\Delta P_x}(f) \tag{2.29}$$

$$S_{\eta,\frac{d\eta}{dx}}(f) = \frac{1}{\Delta x}\left[\frac{\cosh kd}{\cosh k(d+z)}\right]^2 S_{P,\Delta P_x}(f) \tag{2.30}$$

where ΔP_x is subsurface pressure difference between two sensors on the x-axis and Δx is differential distance along the same axis. Similarly, pressure power spectra and co and quad-spectra can be computed along y-axis. Once the power spectra, co, and quad-spectra are obtained, the directional spectral density may be determined using the same technique outlined earlier for deep water directional spectrum.

III. DATA ACQUISITION AND SENSORS

A. LOCATION AND TIME

The area of Monterey Bay was selected as the study area which poses complex problems due to the highly irregular bathymetry of the Monterey submarine canyon. Three sensors were utilized for acquisition of wave data. The locations of the sensors are as follows (see Figure 3.1 for bathymetry and location).

No.	Type of Sensor	Location	Depth (M)
1.	NDBC roll/pitch wave buoy	36°48'00''N 122°23'59''W	2003
2.	Array of pressure gauges (4) Marina	36°42.0'N 121°48.9'W	15.0
3.	Array of pressure gauges (4) Santa Cruz	36°57.0'N 122°00.2'W	13.1

Selected wave cases from January 1988 are used in the analysis. During the winter season, weather conditions are generally favorable for creating higher energy waves. In addition, an abnormal storm was encountered in the region. To select a particular date and time, the conditions for narrow band wave spectrum were emphasized to simplify the interpretations.

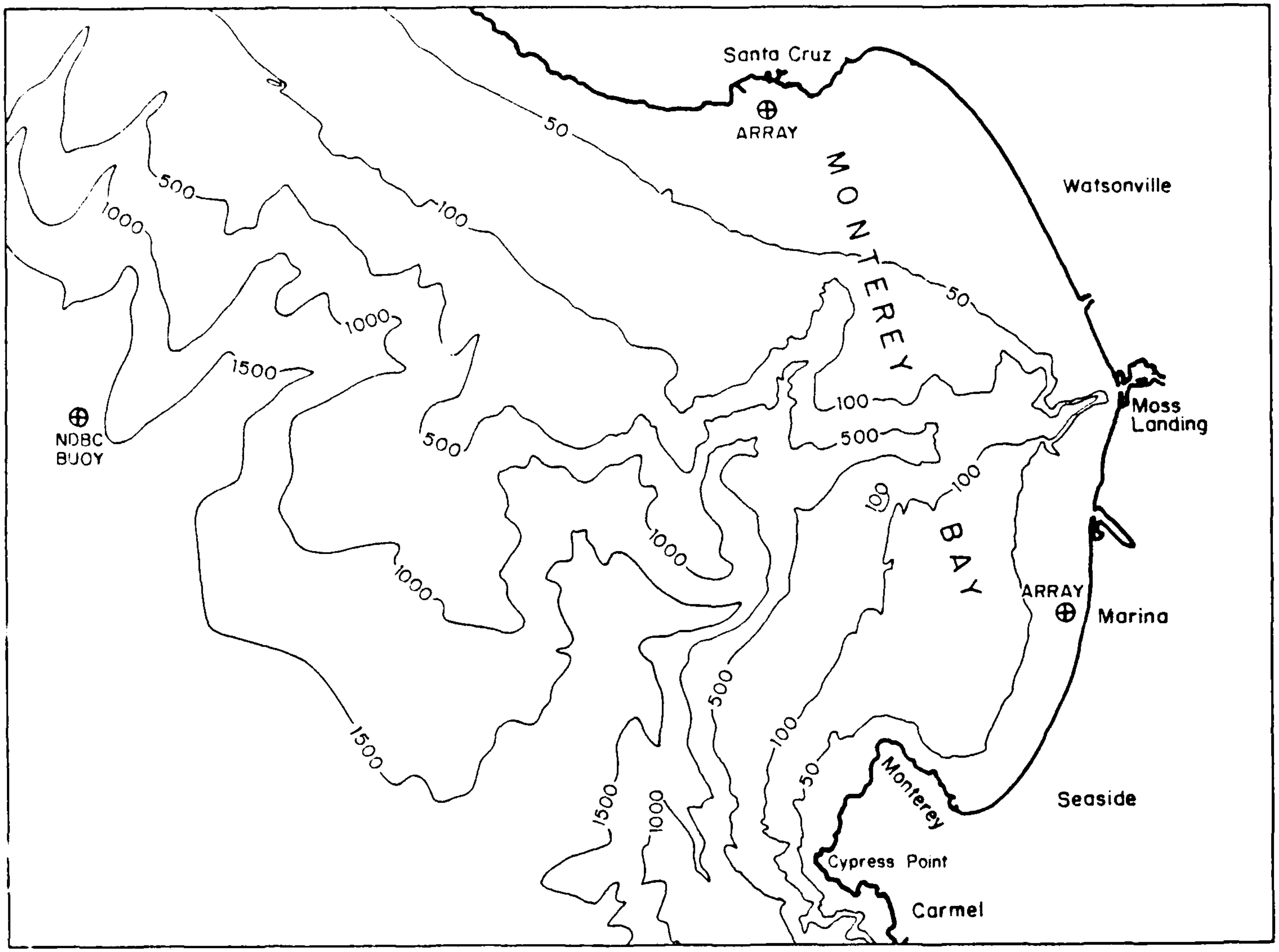


Figure 3.1 Location of Sensors and Bathymetry in Monterey Bay

In comparing coincident data sets, on the average a time lag of one hour was assumed for the waves to travel from the deep water buoy to the shallow water array sites.

B. DEEP WATER WAVE DATA

Deep water wave data for the pitch/roll buoy were obtained from the NDBC. The NDBC 3-meter buoy measures heave, pitch and roll movements with respect to 3-axes magnetometer. The magnetometer axes are aligned with fore and after (bow positive), athwartship (starboard positive) and vertical (positive upward) axes of a buoy. The analog sensor measures vertical acceleration and displacement in addition to hull pitch and roll. The magnetometer bow and starboard output, together with vertical displacement, pitch and roll output from analog sensors are sent to the Directional Wave Data Analyzer (DWDA) after voltage amplification and filtering.

The hull azimuth angle is calculated from bow and starboard magnetic components with necessary corrections made for pitch, roll and hull magnetic fields. The magnetic azimuth is converted into true azimuth using the magnetic variation for the site. The true azimuth is then used to calculate East-West and North-South components of hull slope. One hertz time record of hull displacement and slopes are stored in DWDA for spectral analysis and transmission to shore station through GOES (Steele et al., 1985).

Data provided on magnetic tape included environmental parameters, wave spectra, directional wave parameters and co and quad spectra. Directional wave parameter data were utilized to determine directional wave spectrum. The unimodality of the directional waves was confirmed by studying the regional weather maps to conveniently apply cosine angle distribution function. Two-dimensional spectra are shown in Figure 3.2.

C. SHALLOW WATER WAVE DATA

The calibrated data for arrays of pressure gauges were obtained from the Scripps Institute of Oceanography. The array consists of four pressure gauges located in a six-meter square area which measures pressure in centimeters of water. Each sensor is located about 50 to 100 cm above the bottom. Pressure power spectrum is calculated which in turn is transformed into surface elevation and surface slopes $_x$ and $_y$, spectra by applying linear wave theory transfer function as described in Chapter II.

For the computation of the cross-spectra, the following values were utilized:

- sampling interval = Δt = 2 sec,
- total data points = 2048,
- data points for one sample = 128,
- total record length = 32 min,
- Nyquist frequency = 0.25 Hz

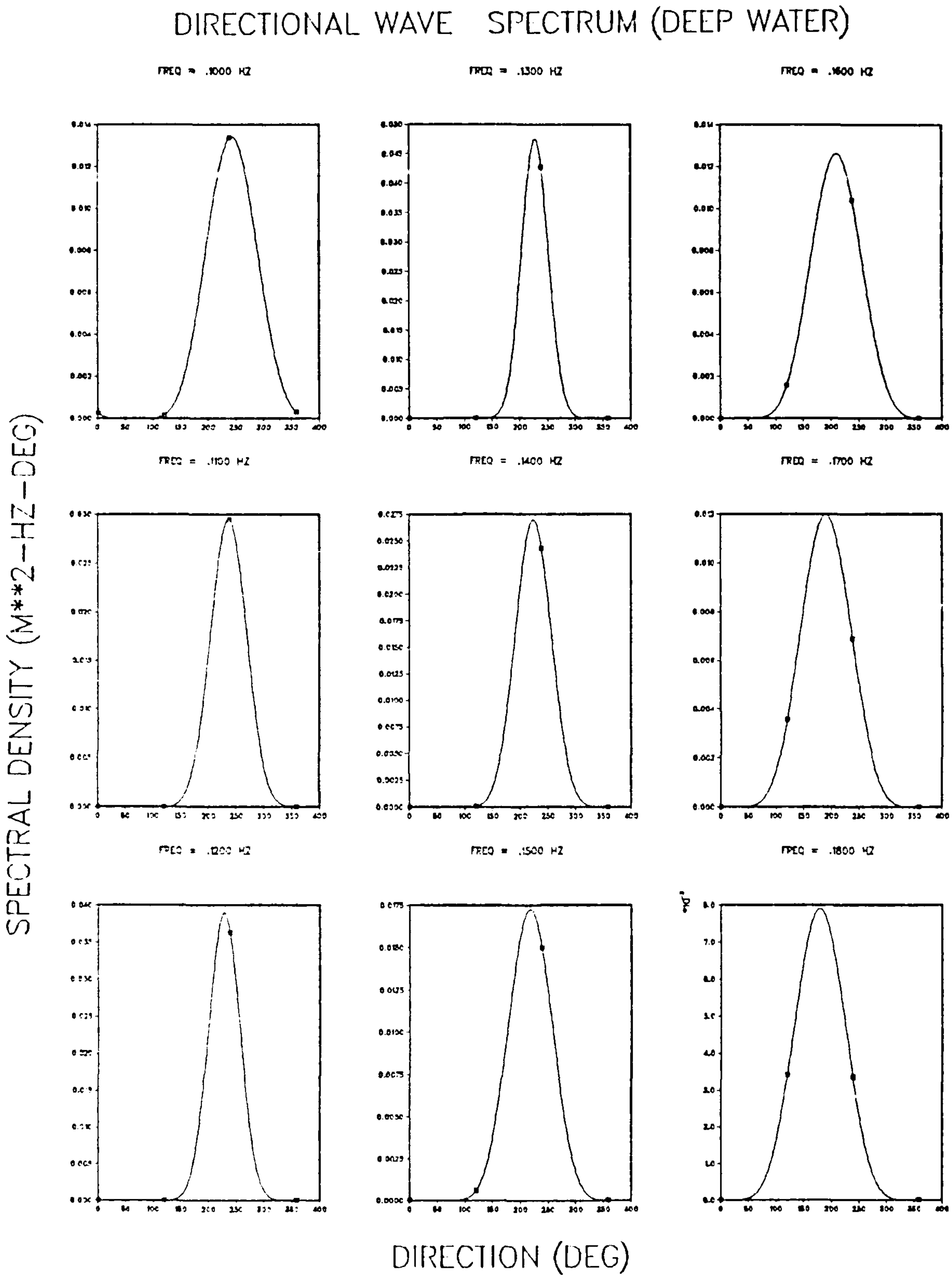


Figure 3.2 Part of Two-dimensional Deep Water Wave Spectrum, 3 January 1988, 0800 PST

- frequency bandwidth = Δf = .00391 Hz,
- degrees of freedom = ν = 32.

The spectra were cut at frequency of 0.2 Hz in the higher frequency range due to spurious overestimation of the transfer function. The wave period for high quality estimate is approximately 5-20 seconds.

To ensure accuracy and to check computations, the results were compared and verified with the published data in the Coastal Data Information Program of January, 1988. Two-dimensional spectrum at the peak energy frequency is depicted in Figure 3.3 for the Marina array on 29 January 1988.

D. OCEAN WAVE TRANSFORMATION PARAMETERS

The Dobson linear refraction model has been used to estimate the shoaling transformation of the wave directional spectrum. The reverse projection method is employed. That is, the refraction program was run backward from shallow to deep water, commencing at a shallow water location in required depth, for Marina 15 meters and for Santa Cruz 13.1 meters. Rays of a particular frequency were propagated offshore in increments of .2 degrees (0.2°) over the range of all possible incoming wave angles. Once the depth is deep, the rays are stopped and deep water wave directions are recorded. The rays were then returned to retrace the same path to verify the backward refraction and to calculate transfer coefficients K_R, K_S and J.

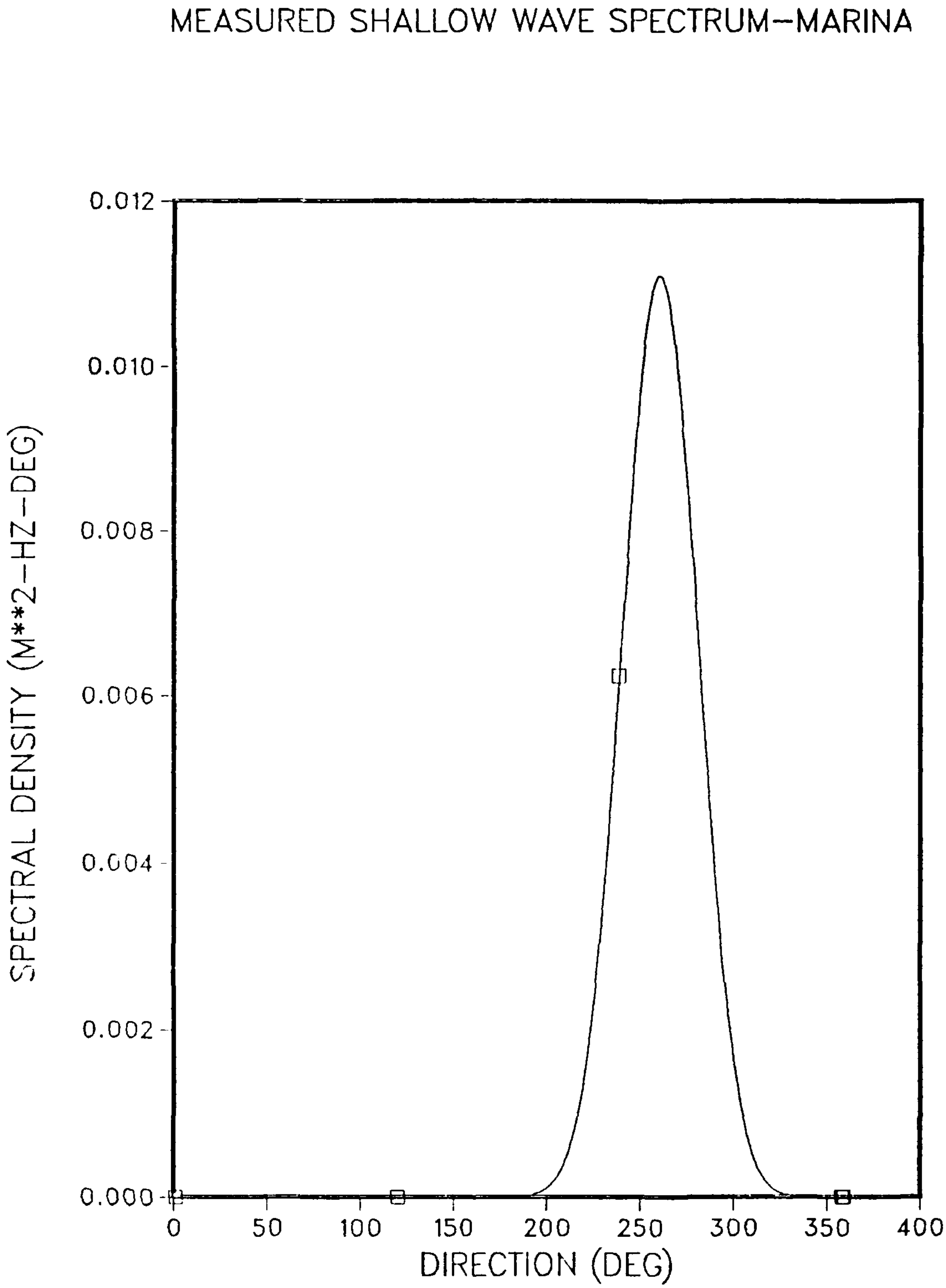


Figure 3.3 Two-dimensional Shallow Water Wave Spectrum Measured at Marina on 29 January 1988

To include refraction over the shelf to the north of Santa Cruz, supplemental bathymetric data were meshed into the model. The shelf was more coarsely gridded to a 1 Km grid because of the large area involved and due to the fact that the waves are primarily in deeper water while traversing the shelf. The transfer parameters calculated from the two areas were multiplied together to determine a close relationship for use in spectral transformation. Some example ray traces are shown in Figure 3.4.

1. Shoaling Coefficient

Shoaling coefficients, K_S^2, computed from the model for Marina and Santa Cruz sites are depicted in Table 3.1.

TABLE 3.1

SHOALING COEFFICIENTS, K_S^2

Freq. (Hz)	Shoaling Coefficients Marina (15 m)	Santa Cruz (13.1 m)
0.030	2.204	2.351
0.040	1.689	1.796
0.050	1.389	1.472
0.060	1.197	1.263
0.070	1.068	1.121
0.080	0.979	1.022
0.090	0.918	0.951
0.100	0.877	0.902
0.110	0.851	0.868
0.120	0.837	0.847
0.130	0.834	0.836
0.140	0.839	0.834
0.150	0.850	0.839
0.160	0.867	0.850
0.170	0.887	0.865

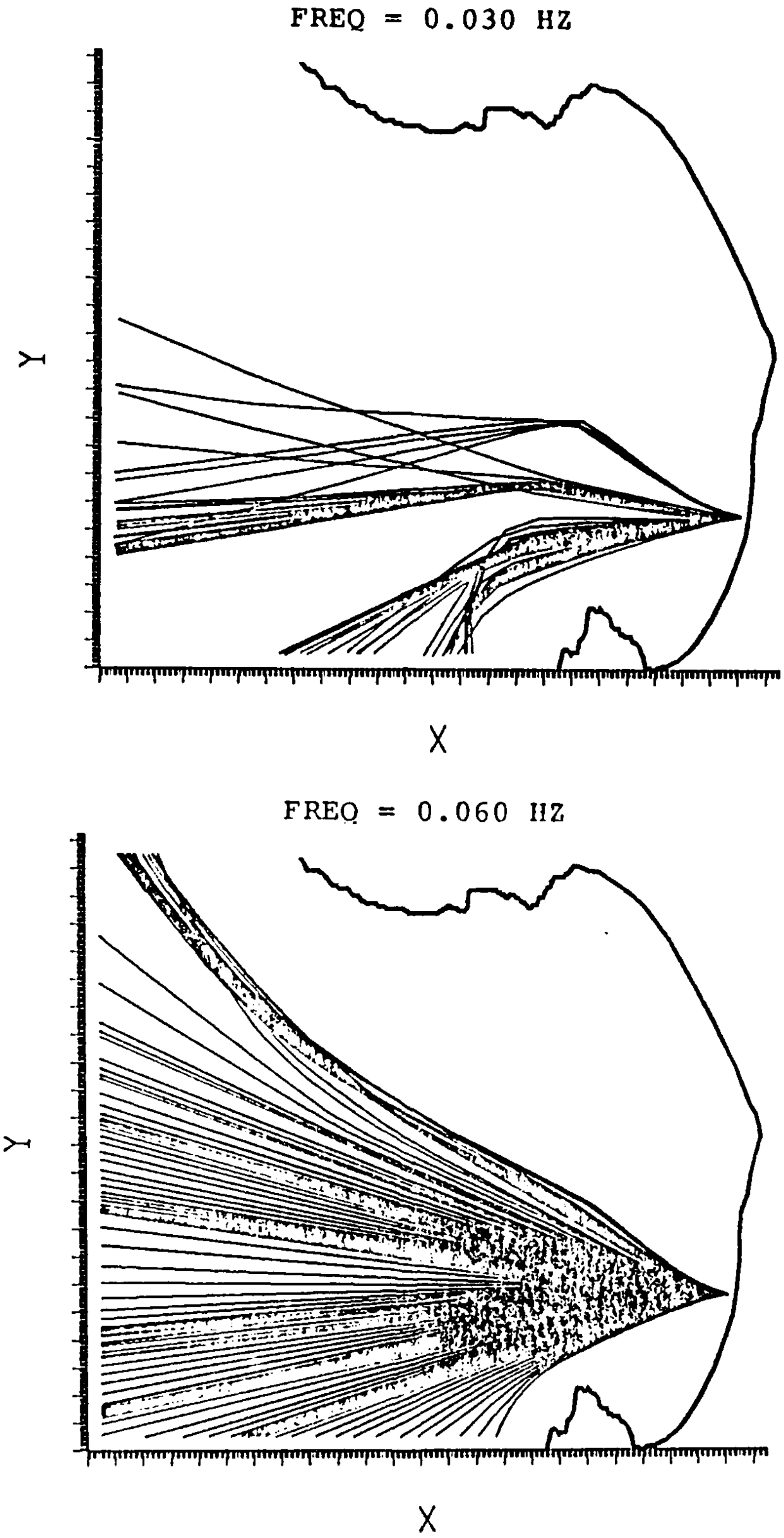
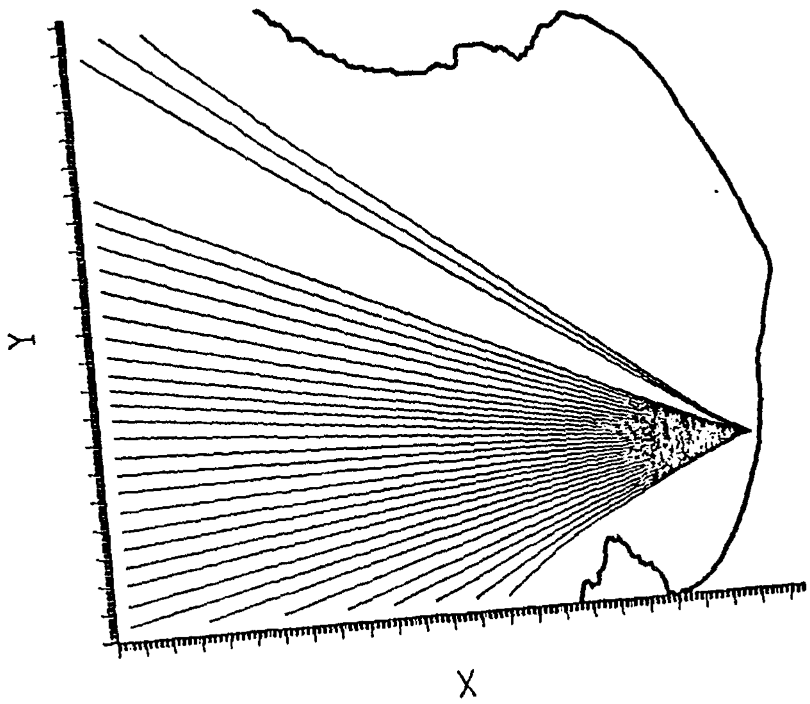


Figure 3.4 Ray Traces from Marina to Deep Water

FREQ = 0.090 HZ



FREQ = 0.130 HZ

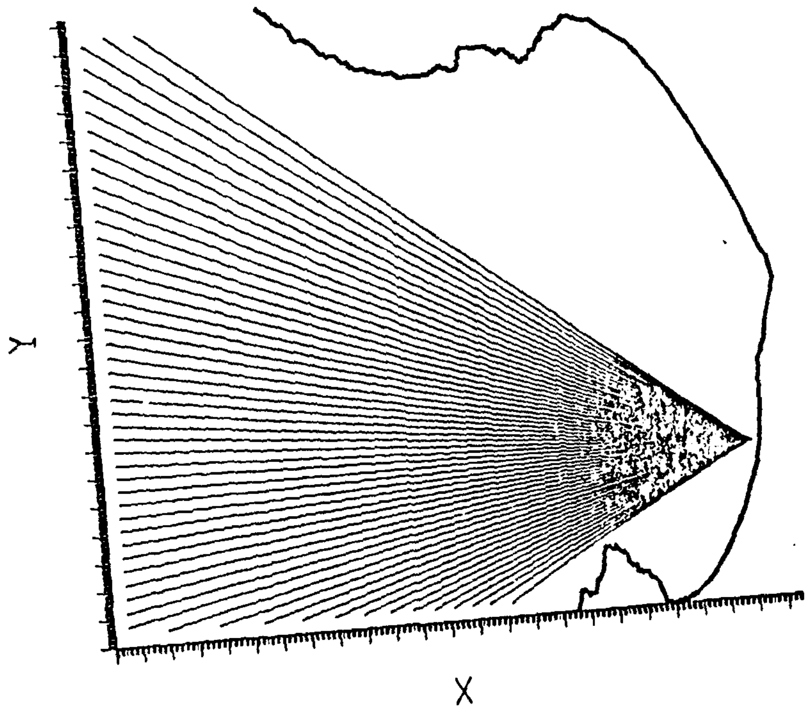


Figure 3.4 (Continued)

2. <u>Refraction Coefficient</u>

The model computes refraction coefficient, K_R, for every 0.2^o of nearshore angle. K_R values were averaged for every degree of deep water direction. In the linear refraction model, an assumption is made that the underwater topography changes are small and gradual but, occasionally, this assumption does not hold in Monterey Bay because of the canyon. As a result, the refraction model sometimes gives erratic estimates of the K_R values. Therefore, all K_R values above three were truncated to the value of three. It is also observed that refraction estimates are more stable in higher frequencies (short period waves) than lower frequencies (long period waves). See Figure 3.5.

3. <u>Jacobian (J)</u>

The Jacobian $(\frac{d\theta_0}{d\theta})$ is calculated from the change in the deep water direction as a function of near shore angles for each wave frequency for all possible wave ray directions. A plot of θ_0 versus θ_S (Figure 3.6) shows a slope of nearly 45^o over most of the angles close to the beach normal. The slope tends to increase as the wave ray undergoes greater refraction. The Jacobian calculated from these values (Figure 3.6) gives similar results, i.e., the Jacobian estimates are constant and closer to one when the approach angle is within 25^o-30^o of the beach normal and increasingly unstable/erratic at extreme limits. Values at the limits may not be useful for computation of directional

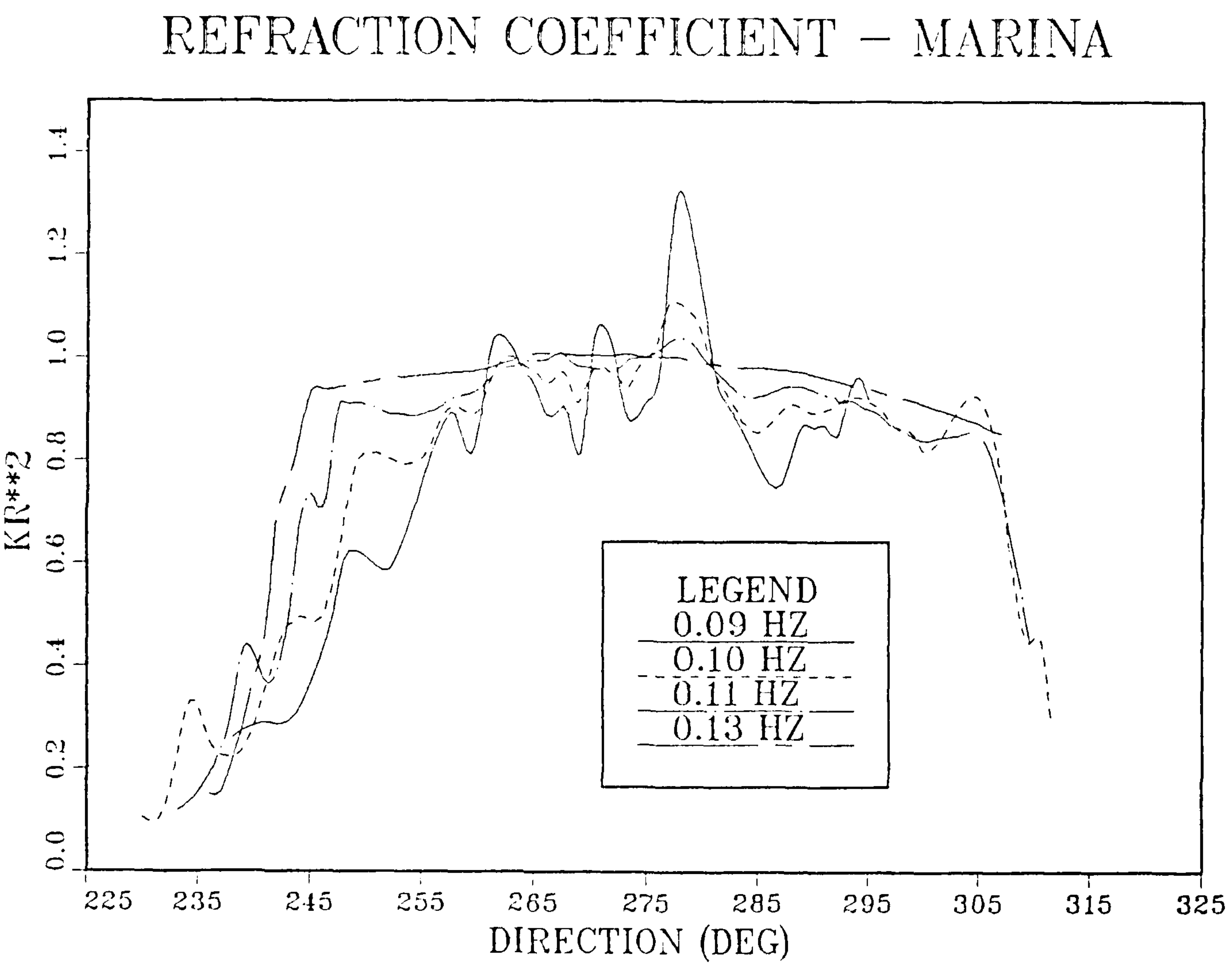


Figure 3.5 Refraction Coefficients for Waves at Marina

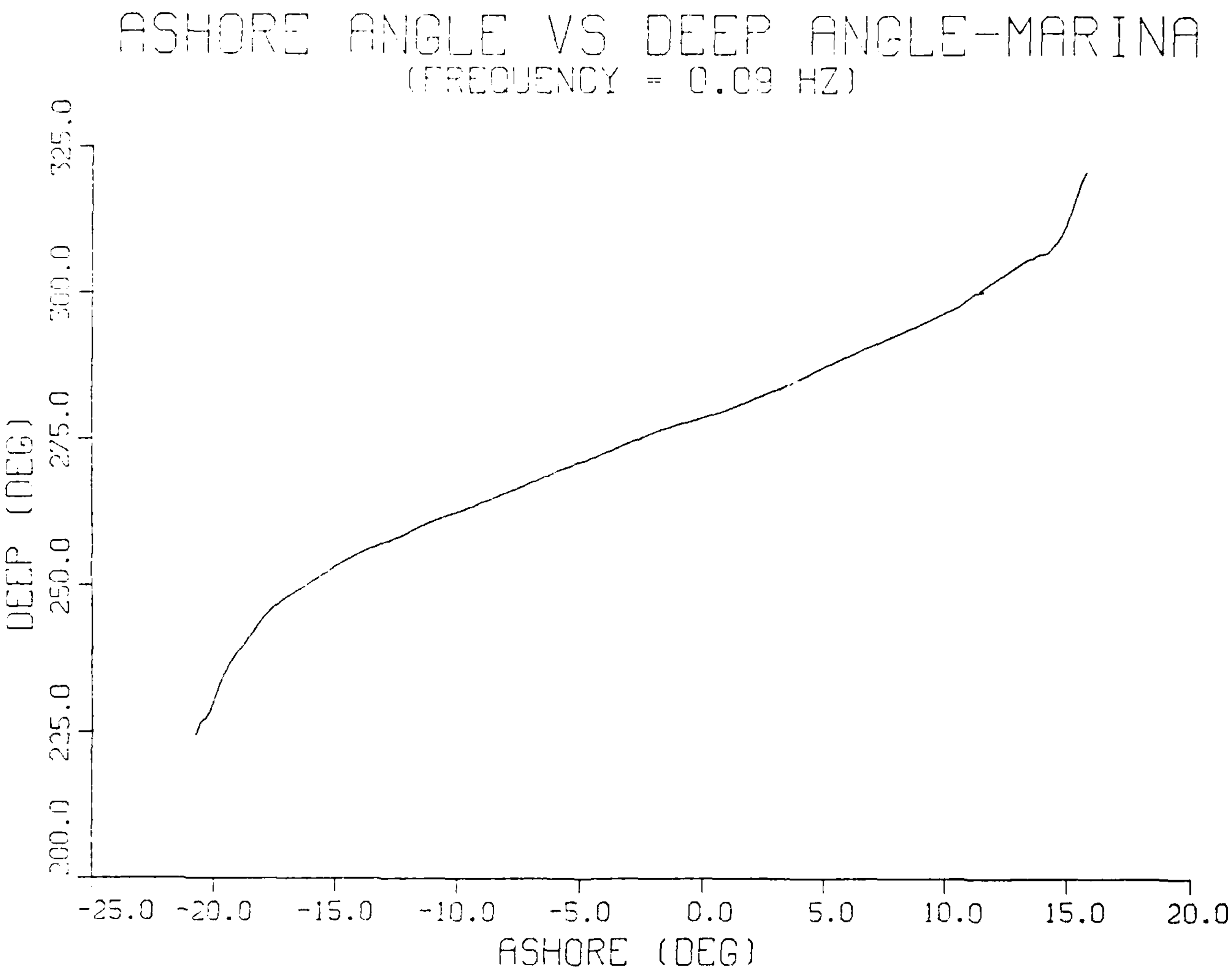


Figure 3.6 Plot of Deep Water Approach Directions vs Ashore Angles for Waves Arriving at Marina Location

spectra particularly in the lower frequency range of .07 Hz and less.

The Jacobian appears to behave similar to the refraction coefficients. The Jacobian estimates at higher frequencies are more stable than at lower frequencies. Figure 3.7 shows the Jacobian estimates at different frequencies for the Marina array location.

4. Transformation of Deep Wave Spectrum

The deep wave spectrum is transformed to shallow water wave spectrum using Equation (2.27). Transformation parameters calculated from the refraction model are multiplied with deep water directional spectrum to obtain shallow wave directional spectrum. Figure 3.8 depicts the transformation of 0.09 Hz wave for different wave directions. The graph of K_S is not shown since at a particular frequency it has a constant value for all directions.

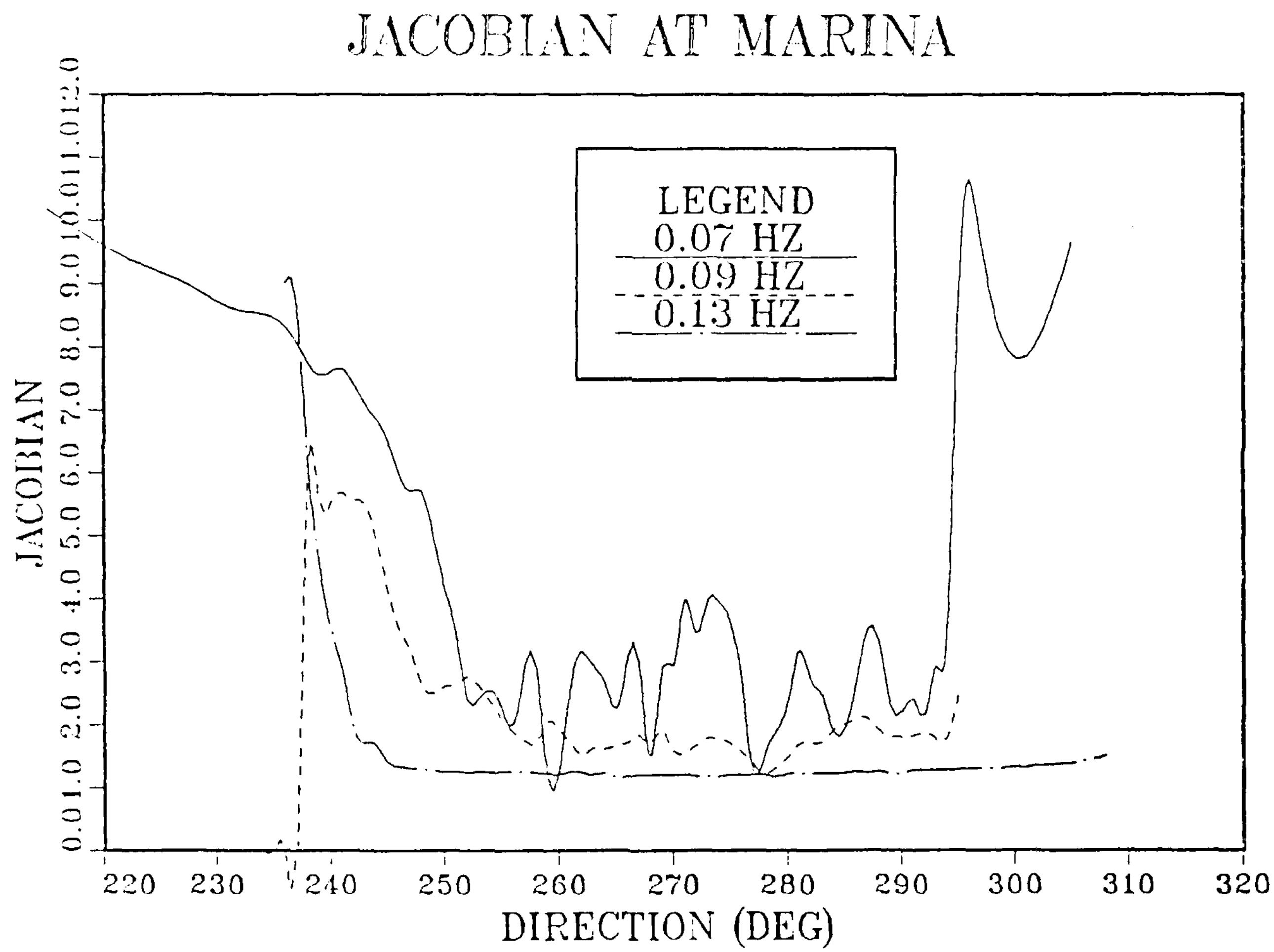


Figure 3.7 Jacobian (J) at Various Frequencies for Marina

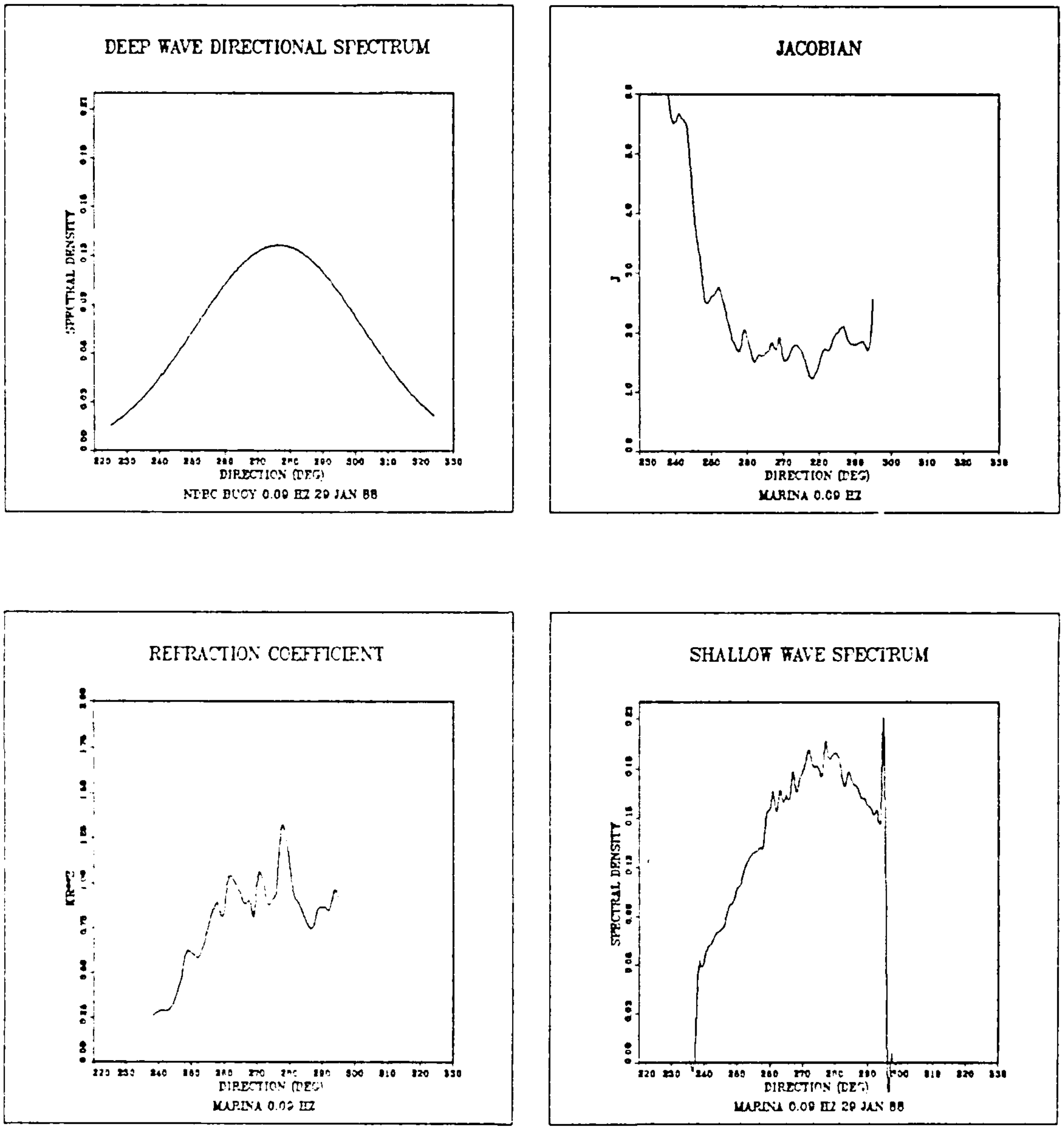


Figure 3.8 Transformation of Spectrum from Deep Water to Shallow Water at Marina for Peak Frequency 0.09 Hz

IV. RESULTS AND DISCUSSION

A. REFRACTION MODEL PREDICTIONS

Transformation parameters, K_R and Jacobian, determined by the linear wave refraction model were variable, with a tendency to increased variability with increased refraction. The lower frequency waves are more refracted than high frequency waves. Therefore, these coefficients tend to vary smoothly in higher frequency (short wave period) and increased in variation with a decrease in frequency. The variability of K_R and J is greater for the Santa Cruz site for which incident waves must undergo significant refraction (see Figures 4.1 and 4.2). The characteristics of each location and refraction model predictions are described here.

The Marina location is shadowed by Point Pino in the Southwest and Point Santa Cruz in the Northwest. High frequency waves, 0.11 Hz and above, are limited to directions from 235°-310° at the array site, whereas low frequency waves may arrive from even southerly directions after considerable refraction (refer to Figure 3.4). Thus, the overall cut-off limits are 235° in the South and 310° in the North. The wave rays, at low frequency, are sparse and tend to form caustics which result in higher variability for estimates of coefficients. The variation of J also depends

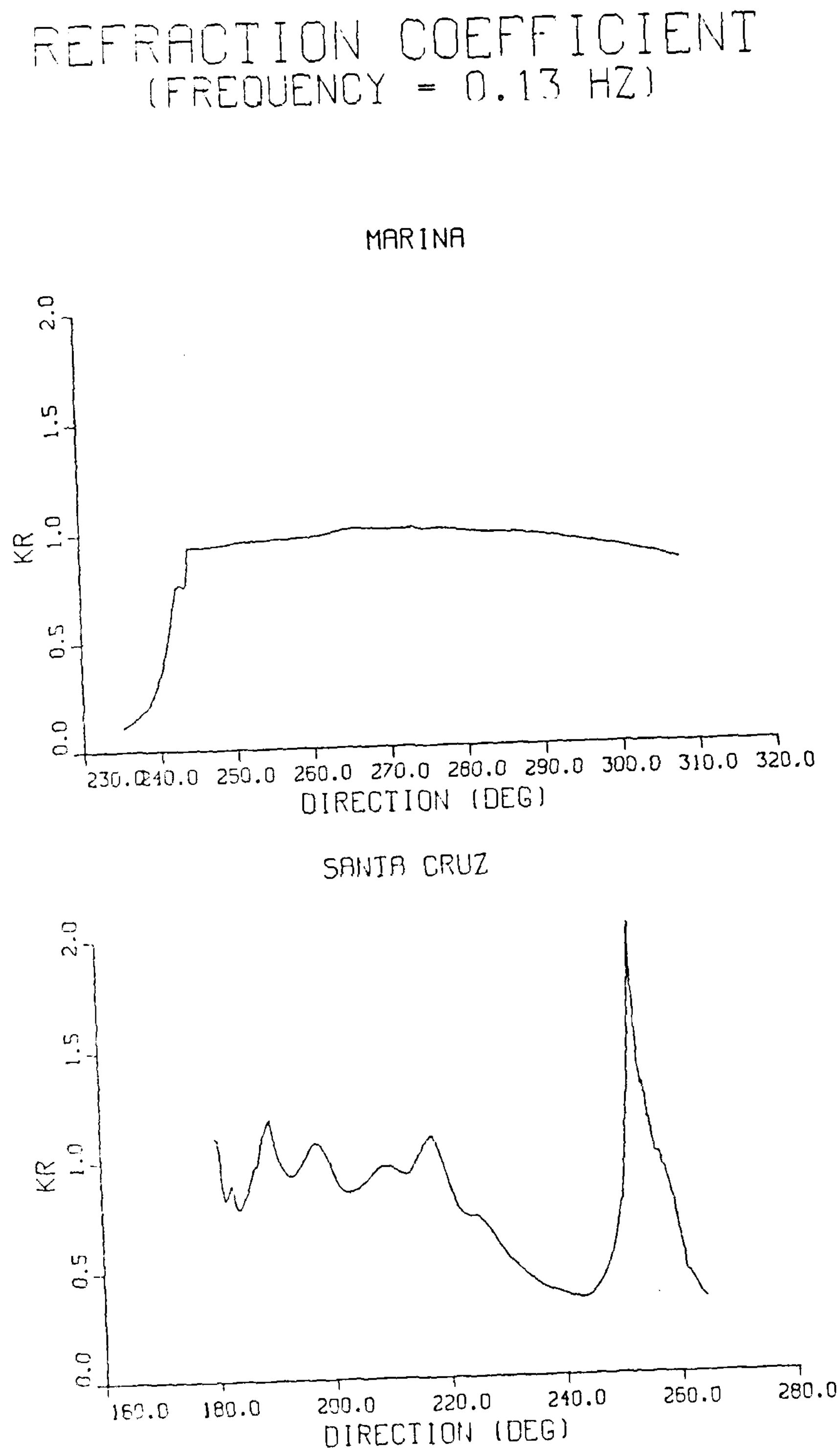


Figure 4.1 Comparison of Refraction Coefficients; Marina and Santa Cruz

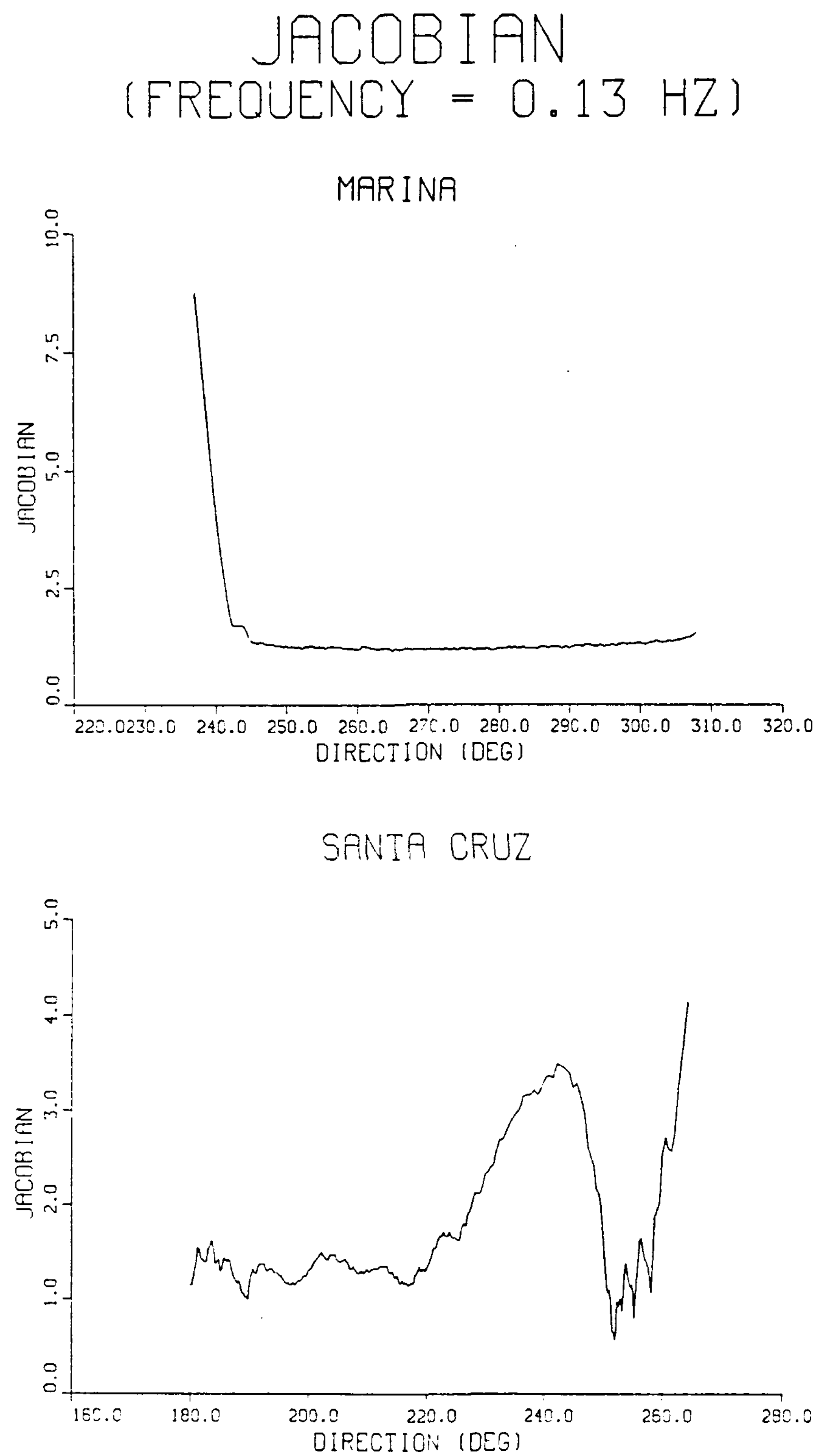


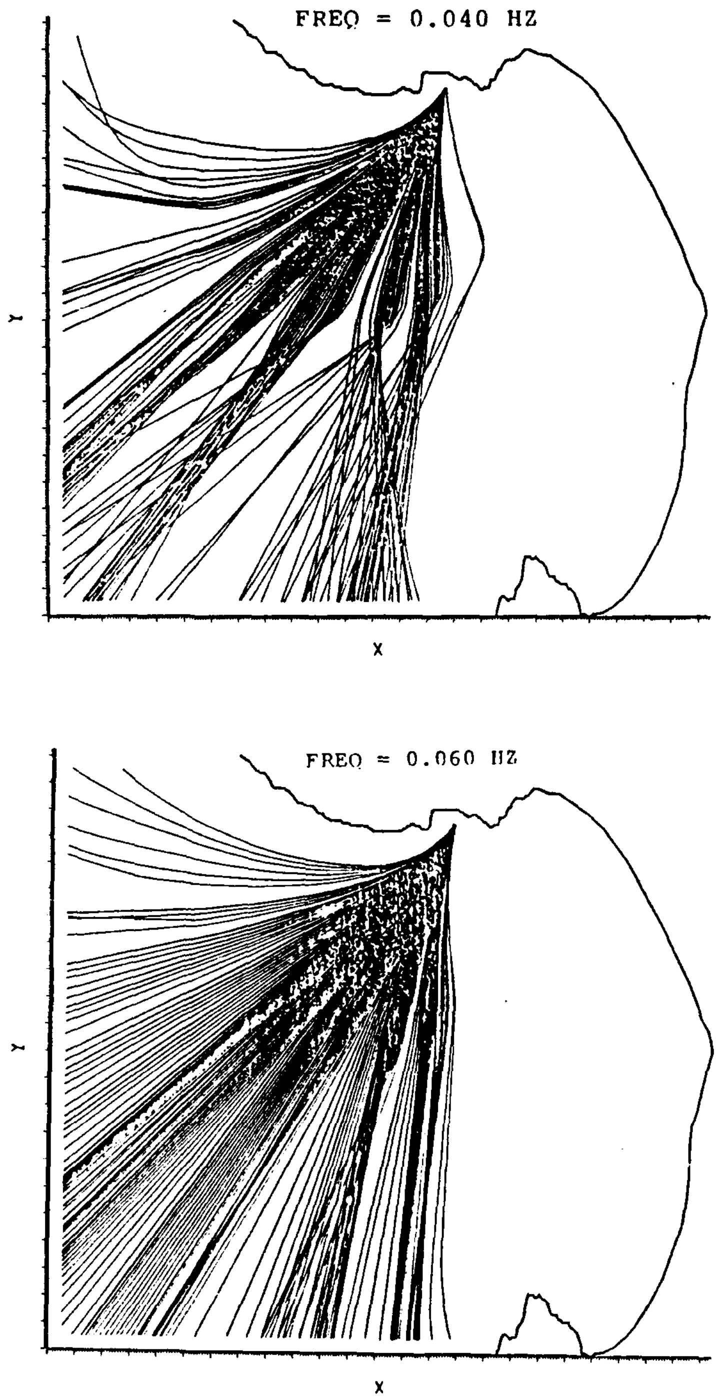
Figure 4.2 Comparison of Jacobian at 0.13 Hz at Marina and Santa Cruz

upon the direction of approaching waves in addition to the frequency. If the wave ray coincides with, or is closer to the beach normal, the J values smoothly vary over the directions. As the angle of approach increases, J tends to vary with increasingly high estimates.

The Santa Cruz site poses more complex problems than Marina because of its location. The waves approaching from the westerly direction makes almost a 90° turn to arrive at the shoreline. The low frequency waves travelling north and north-eastward may arrive at this location. The ray pattern shows (Figure 4.3 a,b) scattered and crossing rays (caustics) with occasional or a few wave rays from the west or higher approach angle at low frequency. This gives very large and highly variable estimates of J. The ray pattern becomes more regular with increase in frequency. It is also noticeable from the ray patterns that the J estimates are generally large and more variable compared with the Marina values (Figure 4.2). The results of K_R estimates behave similarly.

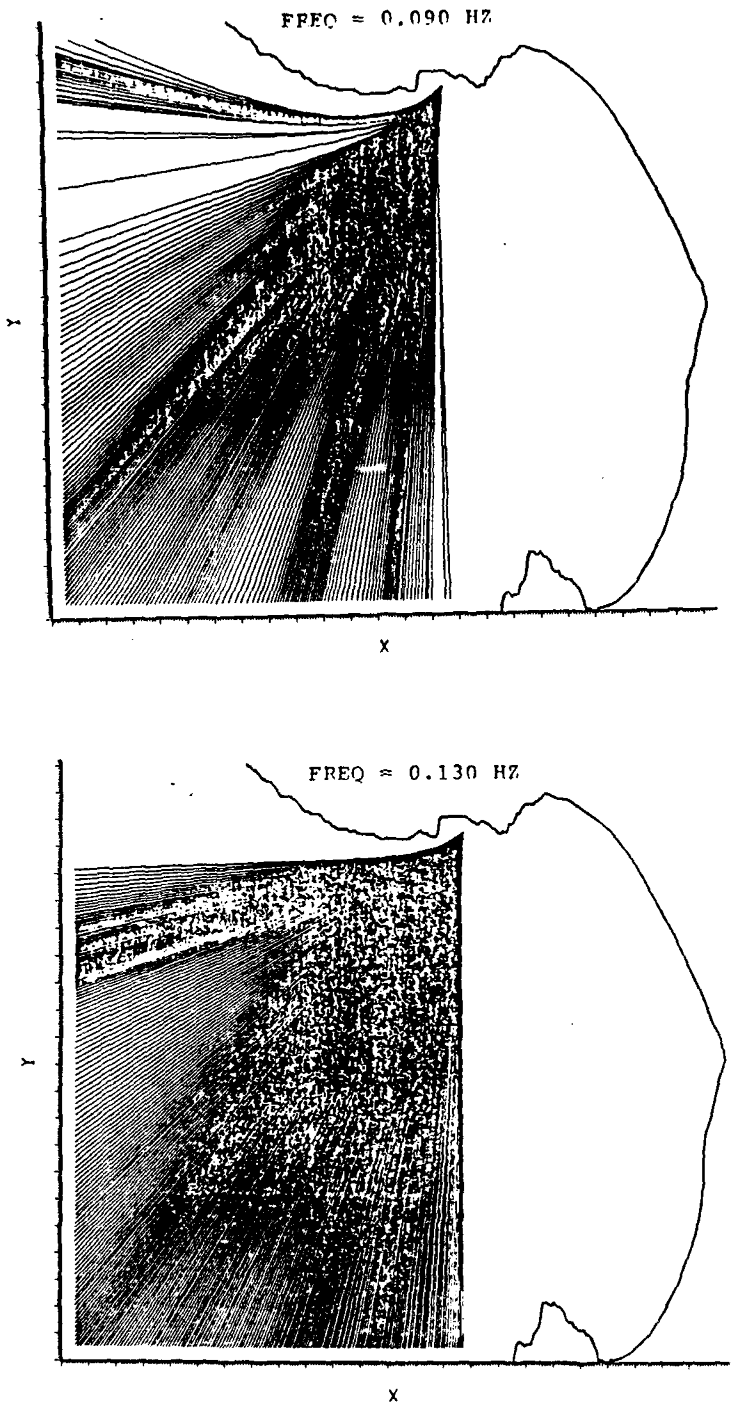
B. WAVE DATA SELECTION

Data for wave measurements for the month of January 1988 were examined. A number of cases were analyzed out of which three are described here. The selected days are 4 January 0800 PST, 18 January at 0200 PST and 29 January at 2000 PST. These periods provide narrow band energy density at different frequencies and direction. Figures 4.4, 4.5, 4.6 and



(a) Low Frequency

Figure 4.3 Wave Ray Traces at Santa Cruz



(b) At High Frequency

Figure 4.3 (CONTINUED)

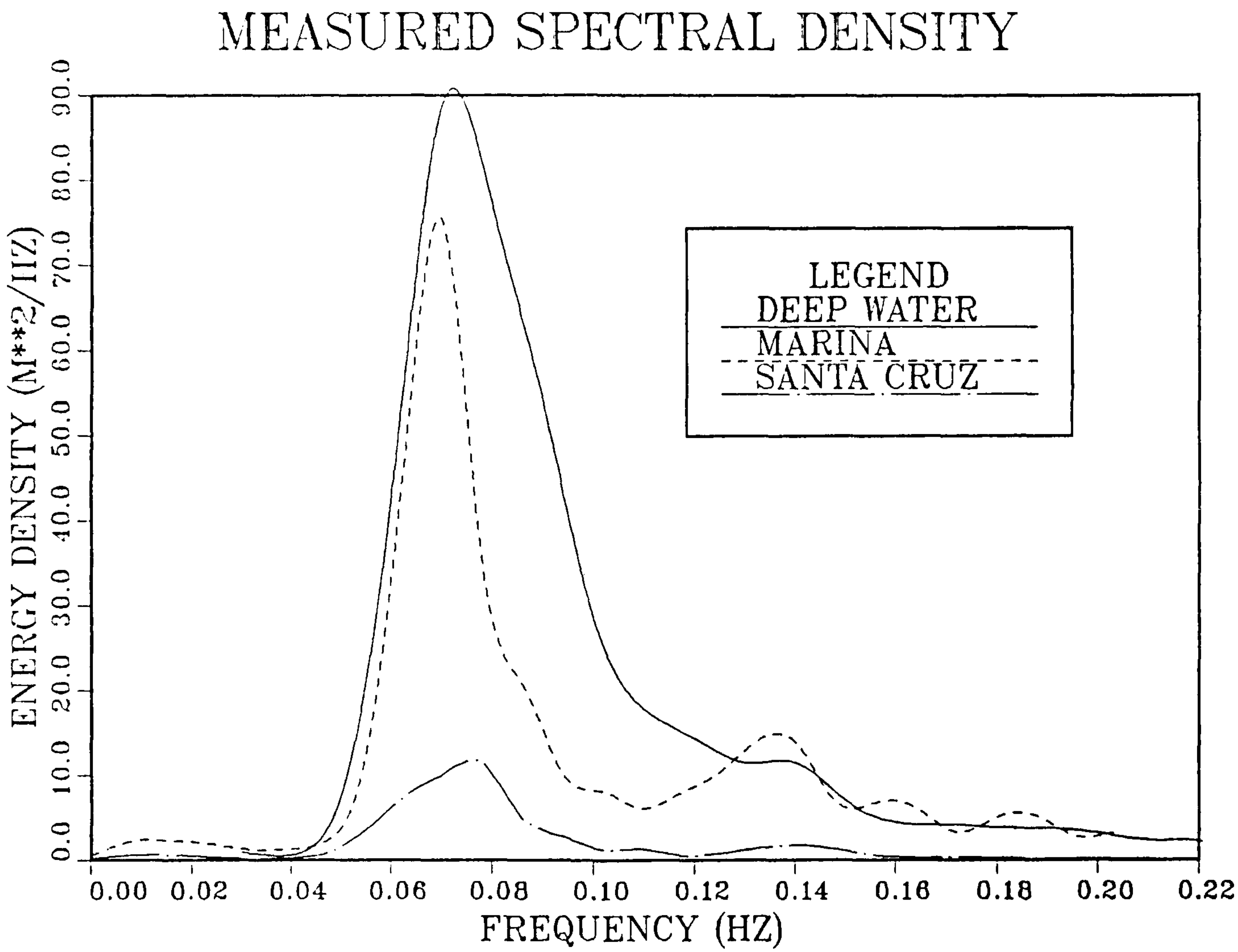


Figure 4.4 Measured Spectral Density on 18 January 1988, 0200 PST

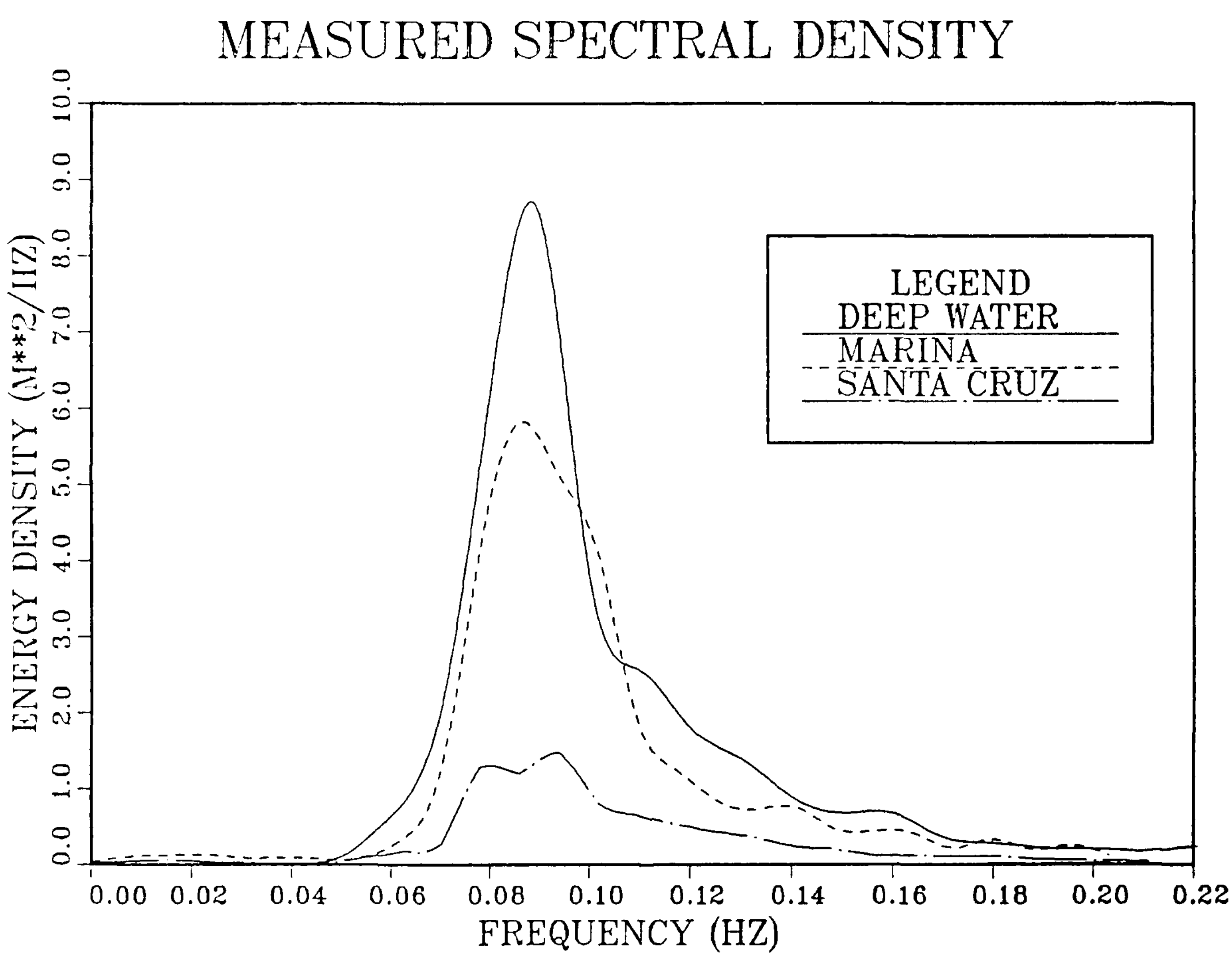


Figure 4.5 Measured Spectral Density for 29 January 1988, 2000 PST

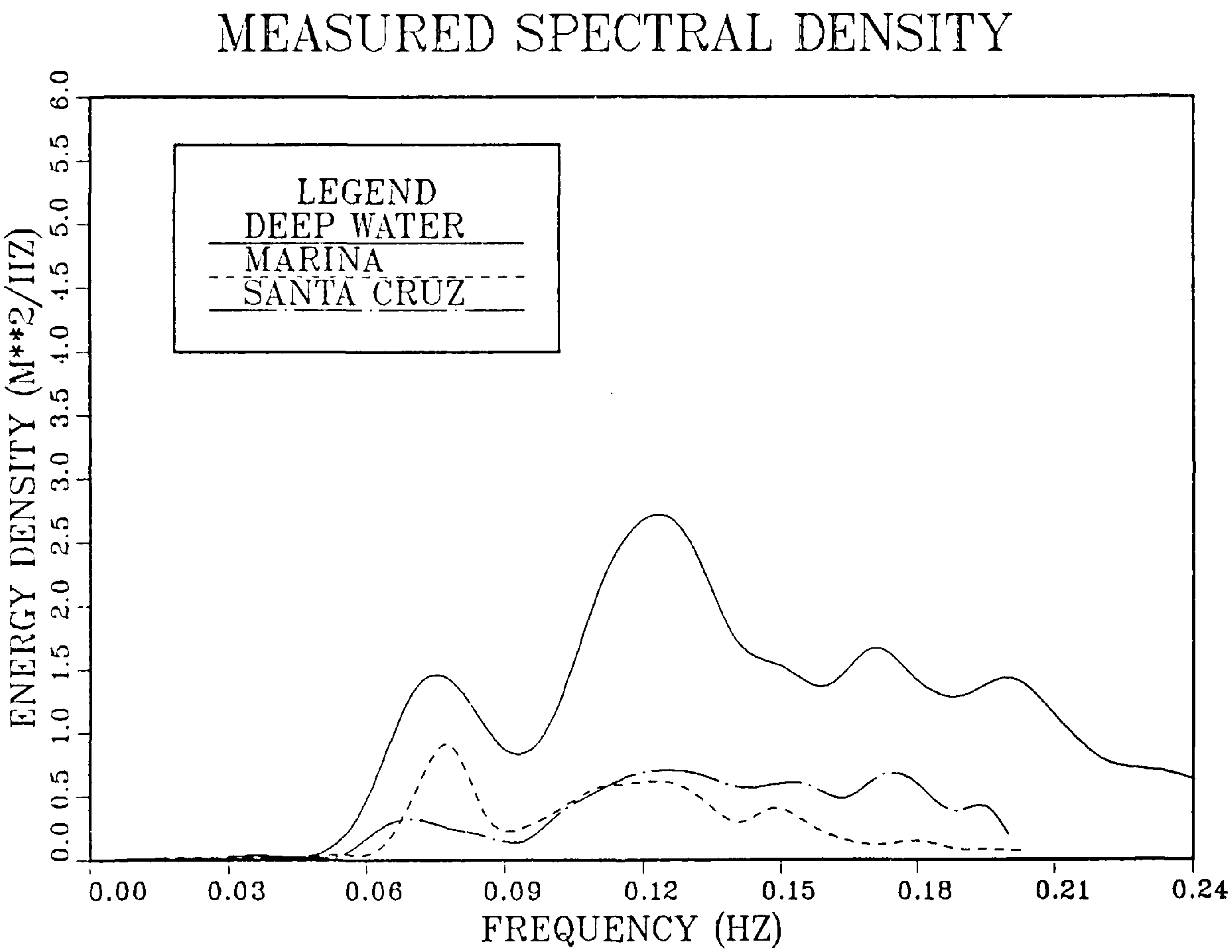


Figure 4.6 Measured Spectral Density for 3 January 1988, 0800 PST

Tables 4.1 and 4.2 provide a comparison of measured directional spectra at the three locations for each day.

TABLE 4.1

DEEP WATER ENERGY DENSITY WITH MEAN DIRECTIONS

FREQ (Hz) (1/sec)	3 JAN 0800 PST DIR (Deg)	ENERGY (m^2-sec)	18 JAN 0200 PST DIR (Deg)	ENERGY (m^2-sec)	29 JAN 2000 PST DIR (Deg)	ENERGY (m^2-sec)
0.0300	173.4	0.000	283.4	0.000	183.3	0.000
0.0400	163.8	0.000	297.3	0.000	234.3	0.000
0.0500	240.5	0.000	265.0	0.000	223.7	0.000
0.0600	214.2	0.000	271.5	26.658	224.1	0.804
0.0700	256.6	1.245	284.6	72.766	256.5	2.125
0.0800	270.0	1.569	292.8	70.196	273.5	6.346
0.0900	268.5	1.198	302.2	38.581	276.6	8.018
0.1000	242.8	1.510	311.8	22.677	277.5	4.256
0.1100	236.7	2.398	308.7	12.638	276.6	3.007
0.1200	228.3	2.755	313.5	11.789	272.2	2.329
0.1300	227.7	2.881	321.3	10.475	268.8	1.570
0.1400	223.9	2.190	313.6	8.089	269.5	1.113
0.1500	217.6	1.699	321.7	3.894	268.7	0.675
0.1600	210.1	1.418	322.5	2.687	265.5	0.609
0.1700	190.5	1.347	312.1	2.315	259.3	0.350
0.1800	179.2	0.889	309.6	1.940	259.6	0.233
0.1900	175.8	0.899	317.2	1.550	267.2	0.184
0.2000	173.6	0.834	311.2	1.147	245.1	0.146
0.2100	165.2	0.670	313.8	0.767	235.6	0.166

TABLE 4.2

WAVE CHARACTERISTICS OF SELECTED CASES

CHARACTERISTICS	3 JAN 88	18 JAN 88	29 JAN 88
$H_{sign.}$	2.2 m	7.9 m	2.3 m
$Freq_{peak}$	0.13 Hz	0.07 Hz	0.09 Hz
$Freq_{mean}$	0.18 Hz	0.10 Hz	0.12 Hz
$Direction_{mean}$	230°	300°	275°

The California coast was hit by an unusually severe storm during 16-18 January 1988 which resulted in high waves, strong winds and sea level well in excess of predicted values (Cayan, et al., 1988). Widespread damage occurred along the Southern California coast. This storm developed about 300 NM west of San Francisco on 16 January 1988 and traveled south-east at about 33 knots. The storm center passed off Monterey Bay on the morning of 17 January 1988 continuing on its SE track and making landfall at Avila Beach in Central California. The bay experienced record high waves 14 hours after the storm passed from the area. A significant wave height of 7.9 meters was recorded on the morning of 18 January 1988. The dominant wave period was about 14 sec with a direction of 284° with narrow band energy measured at deep water and shallow water locations (Figure 4.4).

Deep water energy density with mean direction for selected cases are given in Tables 4.1 and 4.2. On 29 January 1988 measured spectral density of deep water waves and shallow coastal site are shown in Figure 4.5.

Reviewing Tables 4.1 and 4.2 and other published data, deep water waves generally travel eastward during the month of January with slight variability of about 20° on either side of true East. This direction is dominant in the energetic middle frequency range of 0.08 to 0.13 Hz. At other frequencies, the direction tends to be variable.

The third case of 3 January 1988 provides a different view from the normal as it contains sea wave energy at 0.13 Hz frequency with a mean southerly direction of 228° (see Figure 4.6). In this case, deep water waves undergo less refraction to arrive at the Santa Cruz array.

C. COMPARISON OF MEASURED AND CALCULATED SPECTRA

The directional spectra from wave measurement data were determined using a cosine-powered distribution function (Equation (2.26)). The value of the S parameter, in most cases, was less than 20 and relative directional energy was in a broad band of 50° to 60° spread.

Predicted and measured shallow wave directional spectra for Marina are compared in Figures 4.7-4.9. The deep water wave approaching directions are limited between 235° to 310°. These limits change slightly with decrease in wave frequency. In the low frequency range, the predicted spectral energy is overestimated (see Figures 4.7 and 4.8). The deep water spectra of high frequency waves from the West can be transformed into shallow water with reasonaply good estimation comparable with actual wave measurements. The predicted spectrum from westerly waves is shown in Figure 4.9. The predicted mean shallow water direction tended to be more aligned with the deep water spectrum than the measured shallow water spectrum.

The transformation does not appear to appropriately shift the mean direction of spectral peak energy of deep

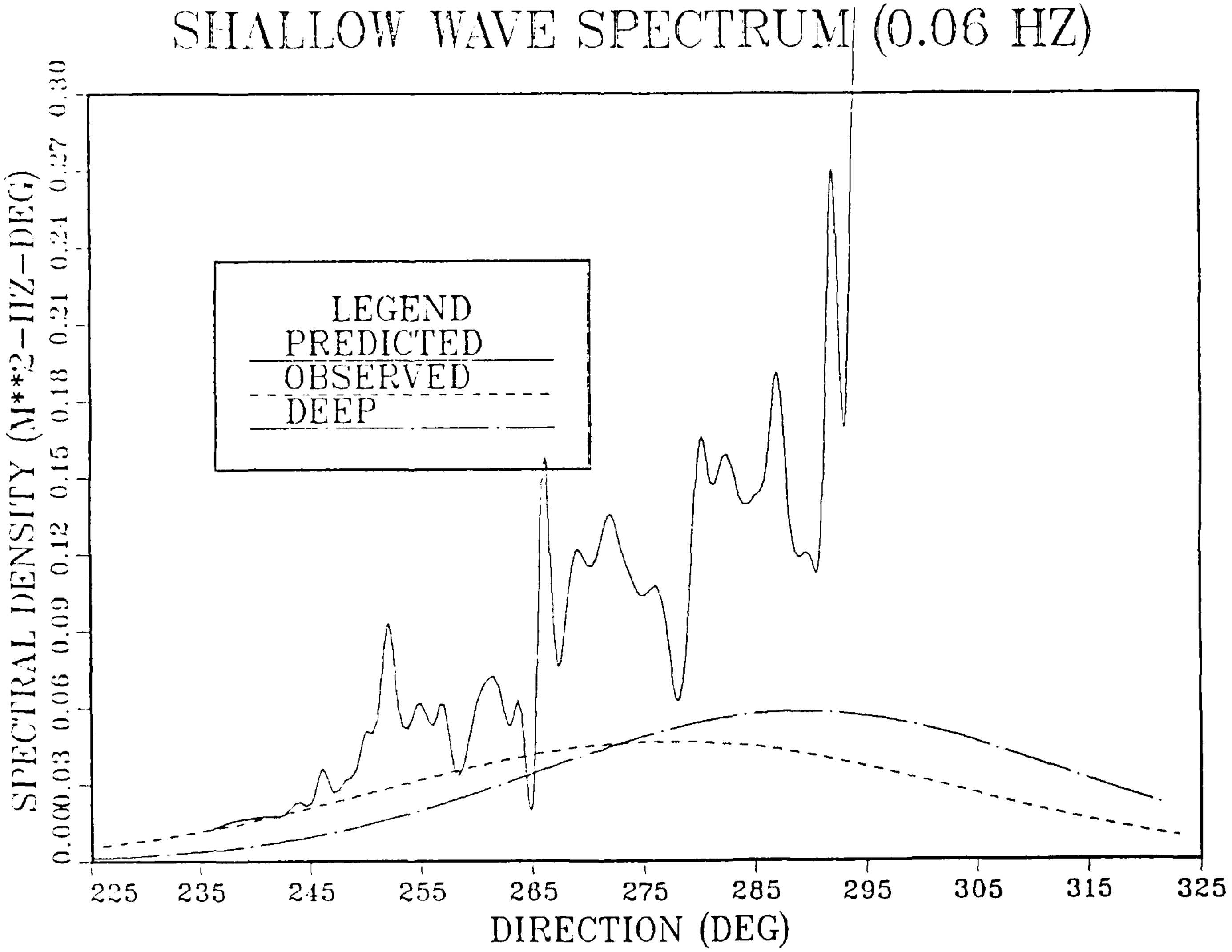


Figure 4.7 Comparison of Shallow Water Spectra at Marina

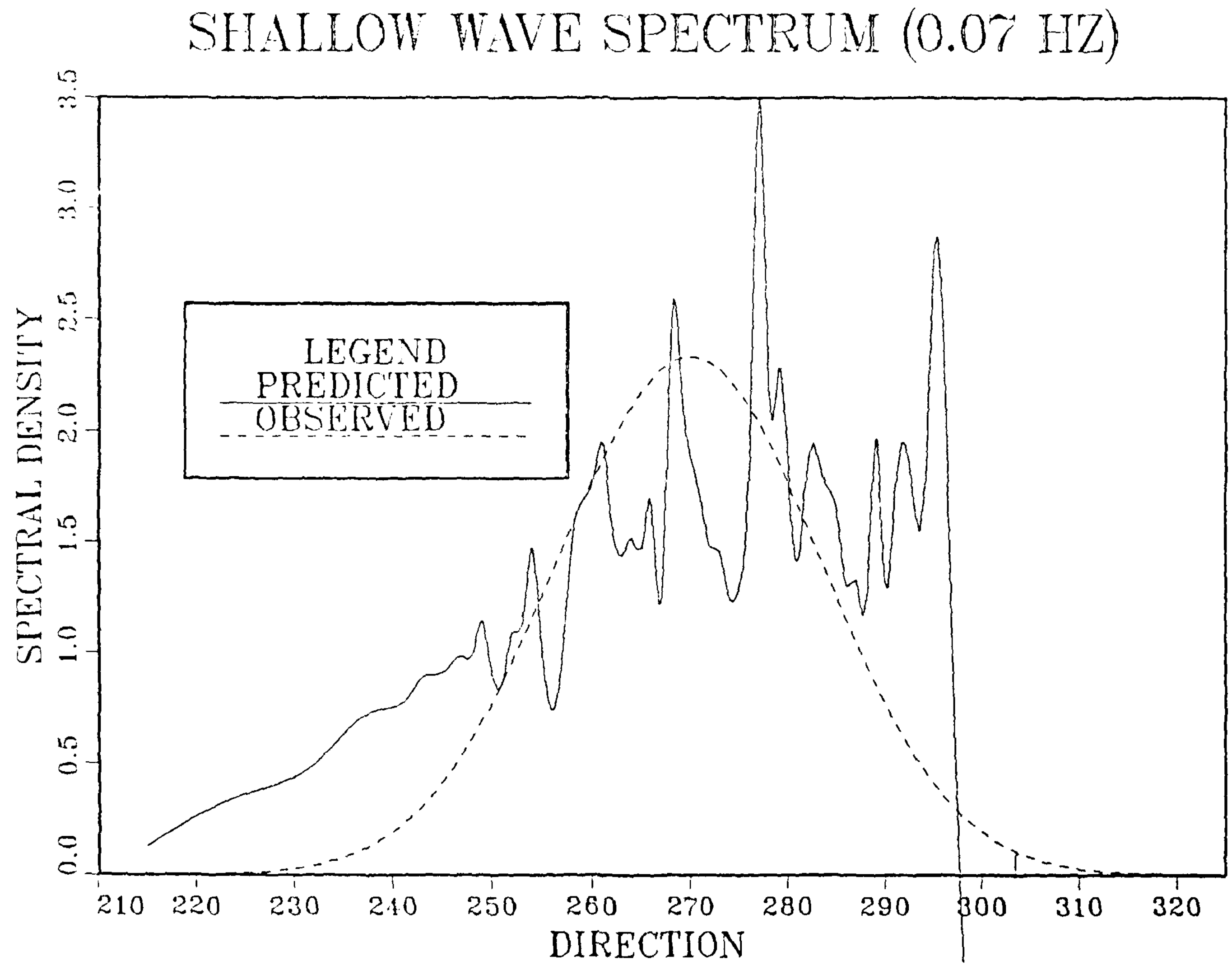


Figure 4.8 Comparison of Shallow Wave Spectra at Marina for 18 January 1988

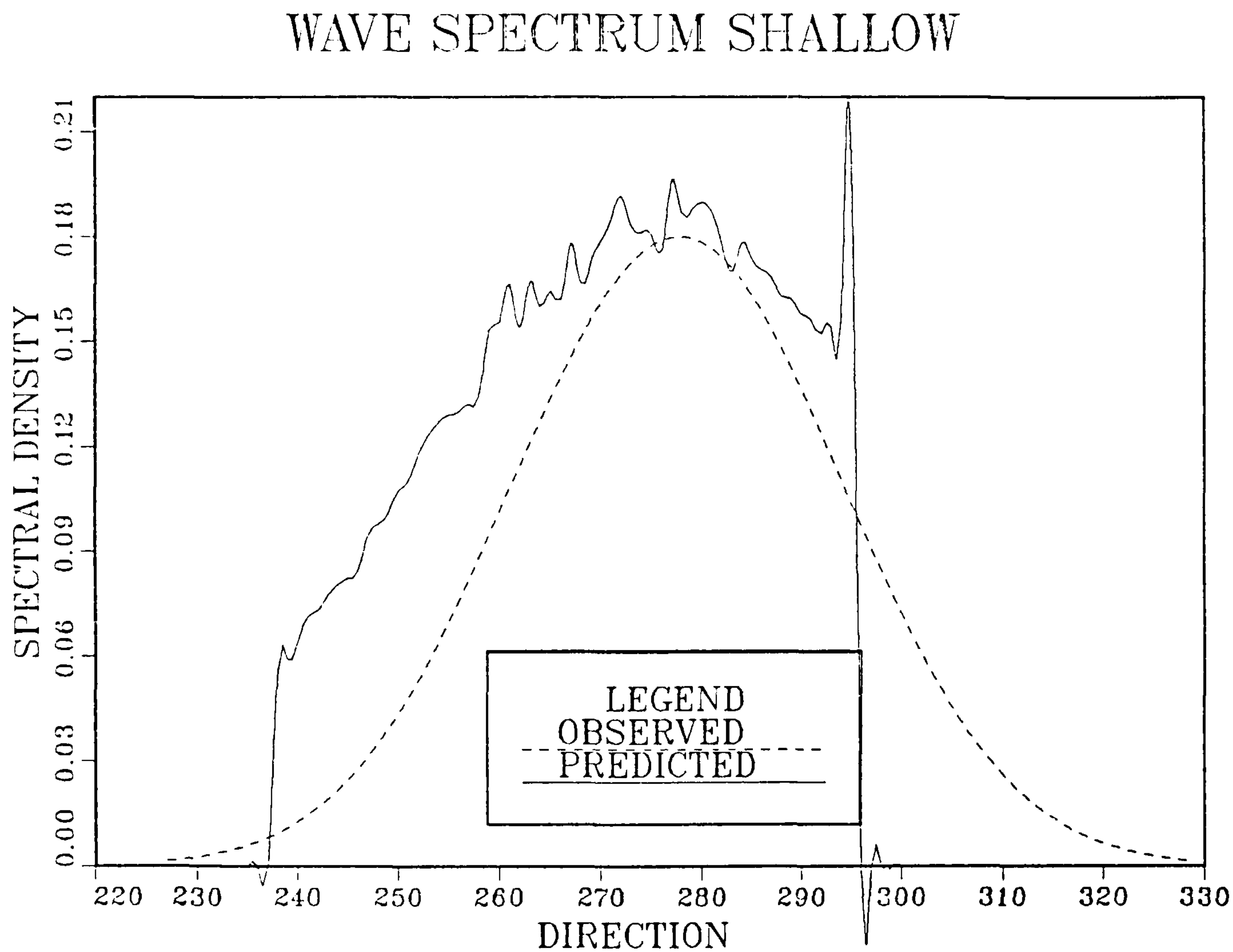


Figure 4.9 Comparison of Predicted and Observed Shallow Wave Spectra at Marina for a Frequency of 0.09 Hz for 29 January 1988

water wave spectra. Two additional cases of 5 January 1988 and 30 January 1988 are shown in Figures 4.10 and 4.11 for high frequency spectra. The mean approach direction for these frequencies in deep water and shallow water are:

		Deep	Shallow
5 January 1988	0.12 Hz	255°	272°
30 January 1988	0.13 Hz	292°	281°

The predicted spectrum in shallow water also indicates high spectral density in the mean direction.

At higher frequency sea waves, the refraction model overpredicts spectral density by less than 18% when summed over all angles at a particular frequency. For swell waves, the results are largely exaggerated.

Results of the Santa Cruz location are depicted in Figures 4.12 and 4.13. The results are quite different from Marina. The predicted spectra are highly overestimated at lower as well as higher frequencies. The relatively simple case of 3 January 1988 was particularly selected for Santa Cruz to consider ocean waves from the southwest direction. Even this simple case does not provide satisfactory results.

These differences are broadly associated with bathymetry. The non-linear effects due to diffraction and currents were neglected which may introduce errors. It is surmised that the complexity of bathymetry in Monterey Bay may contribute significantly to overestimation and variation of transformation parameters.

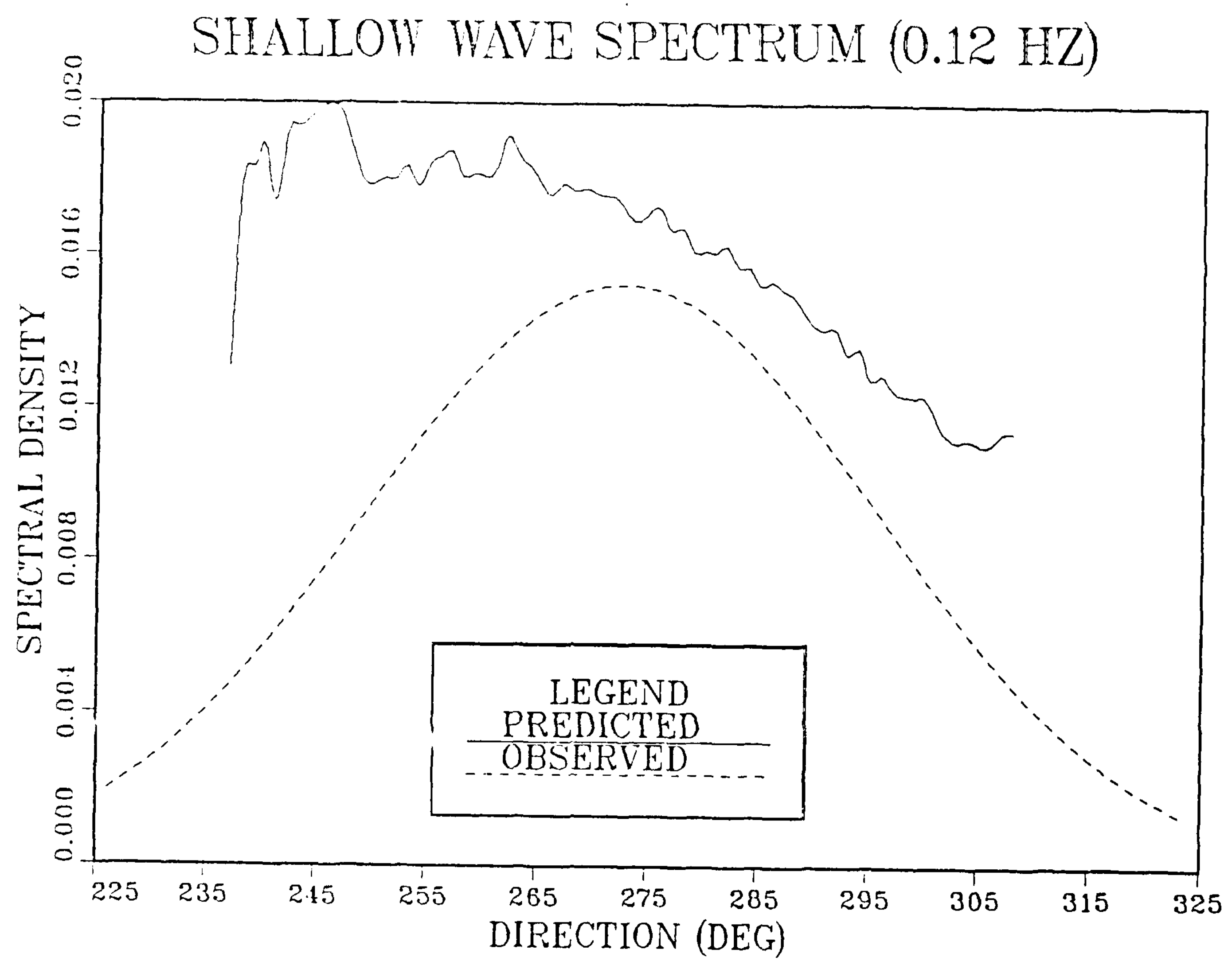


Figure 4.10 Comparison of Predicted and Observed Shallow Wave Spectra at Marina at a Frequency of 0.12 Hz, 5 January 1988

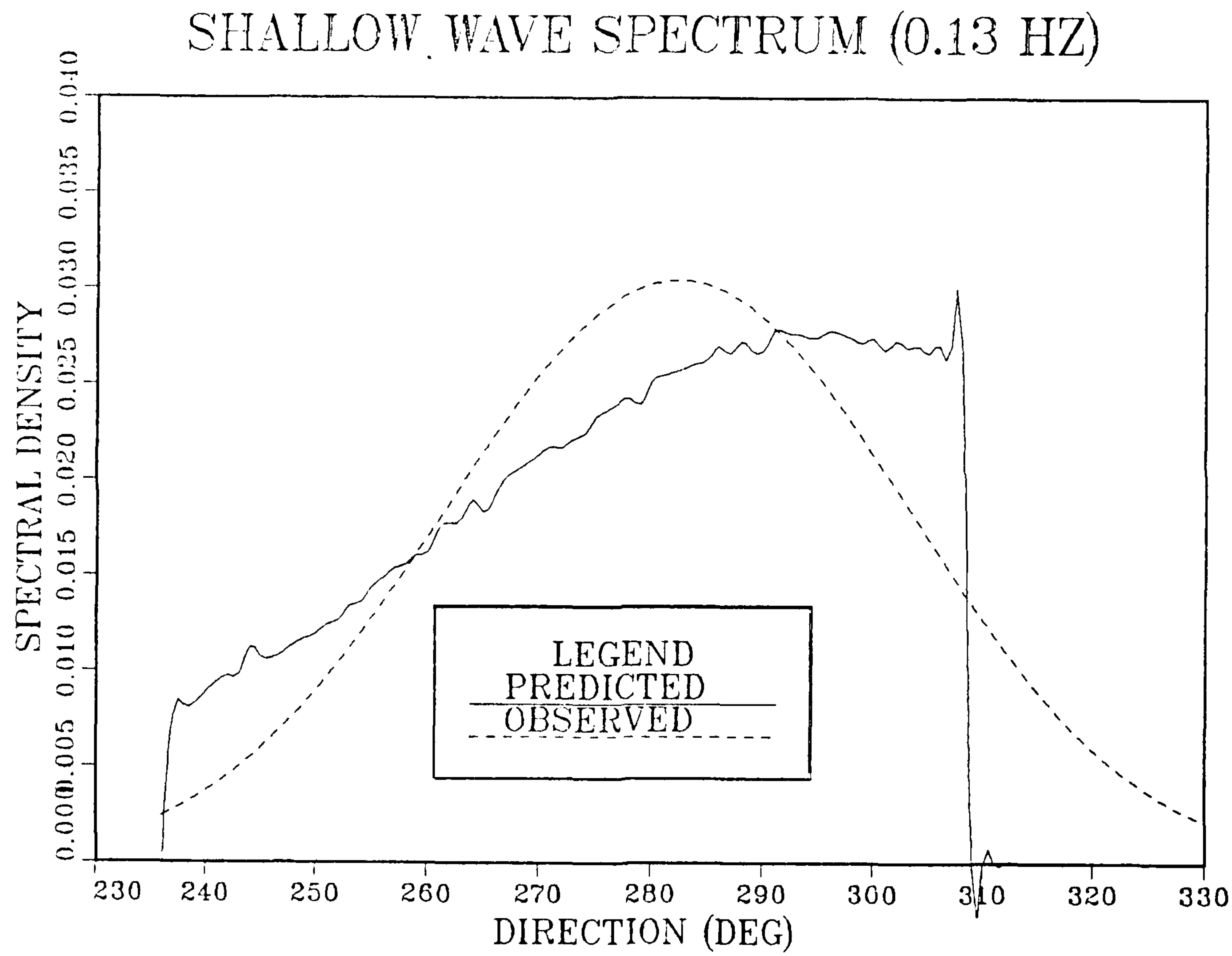


Figure 4.11 Comparison of Shallow Wave Spectra at Marina for 30 January 1988

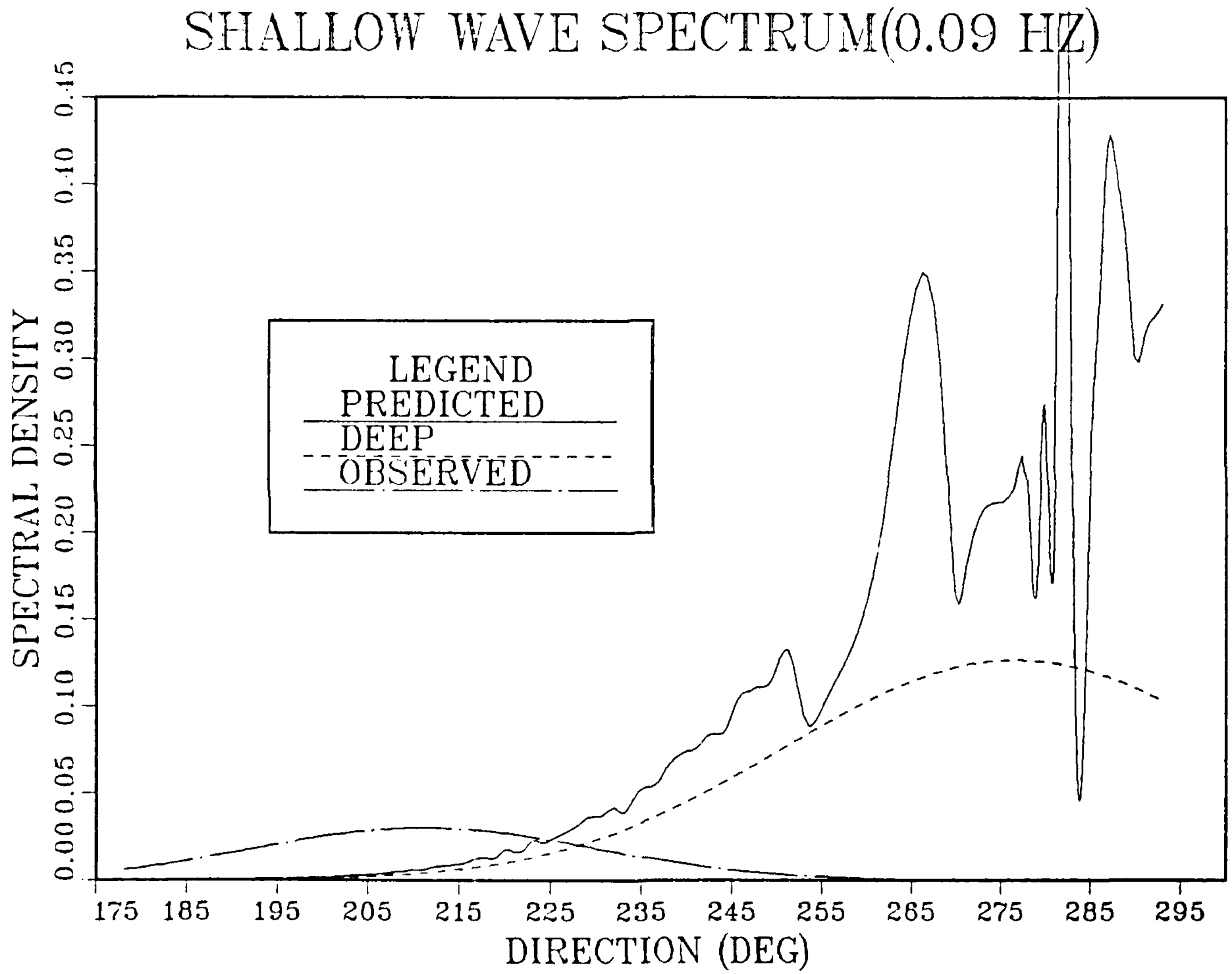


Figure 4.12 Comparison of Shallow Wave Spectra at Santa Cruz for 29 January 1988

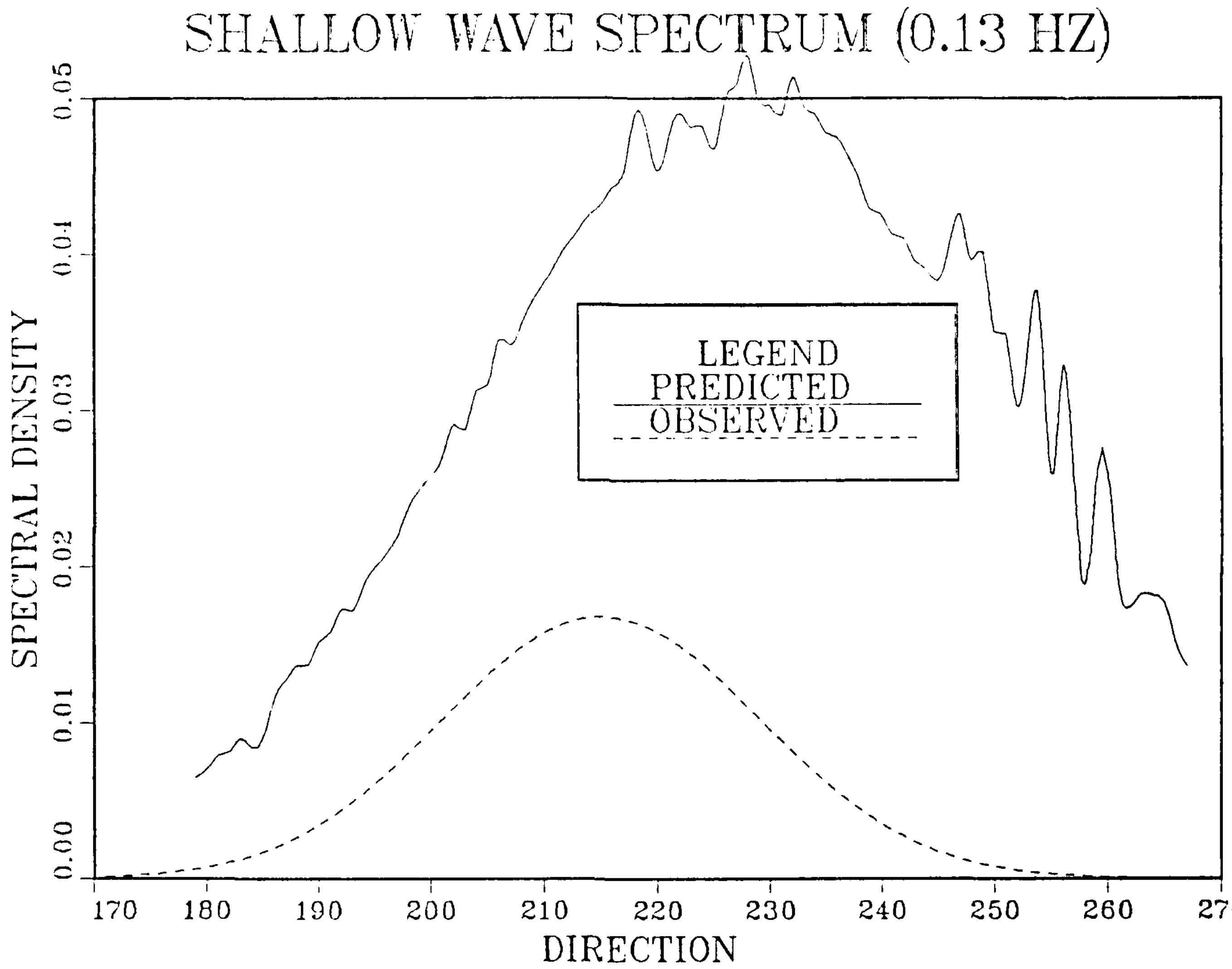


Figure 4.13 Comparison of Shallow Water Wave Spectra at Santa Cruz for 3 January 1988 at 0800 PST

V. CONCLUSION

Deep water directional wave spectrum was determined from data acquired from the NDBC buoy and transformed to shallow water using the linear wave refraction model. The transformed spectrum was compared with measured shallow wave spectra at Marina and Santa Cruz.

The transfer parameters of refraction coefficients and Jacobian predicted by the linear refraction model are highly dependent upon wave period and direction of approaching waves. The Jacobian appears to be more sensitive to wave period and direction than the refraction coefficient. These coefficients are generally stable at higher frequencies and/or when the wave ray is normal or nearly normal to the coastline.

The model tends to overestimate the transformed spectrum. The estimates are less than 18% in excess of the measured spectral density for easterly waves at frequencies of 0.09 Hz and higher for the Marina site. At lower frequencies the results are largely exaggerated.

The model provides unrealistic large variations in wave height calculations for waves at Santa Cruz. This is presumably due to its location, orientation of shoreline with prevailing north-easterly wave direction and the presence of Monterey submarine canyon.

It is clear that the strong irregularities of bathymetry in Monterey Bay causes non-linear effects. The linear refraction model is not suitable for handling such problems. Further modeling efforts including non-linear effects are needed for Monterey Bay wave study.

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